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SEDIMENTS AND EXPONENTIAL CURVES

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ABSTRACT

Several sedimentary situations are discussed in which the suggestion is strong that negative exponential functions are involved. Methods of testing data for these relationships are given in the case of pebble size on a beach, the thickness of a loess deposit, the profile of an alluvial fan, and the dispersion of glacial boulders in a boulder train. In each case the controlling coefficient is determined, and in a summary at the end of the paper are discussed the implications of the exponential function as applied to geological phenomena. It is tentatively suggested that the laws of probability play a part in determining the exponential functions, and that these laws may find a wider application in the study of sedimentary deposits.

INTRODUCTION

An exponential function may be defined as one in which the dependent variable, y, increases or decreases in geometrical progression as the independent variable, x, increases in arithmetic progression. There are many phenomena in the natural sciences which obey exponential laws; in physics and chemistry they are common, but as yet they have been applied to only a relatively few situations in geology.[1] Among sedimentary phenomena there appears to be a wide field for the use of exponential functions, and, moreover, the critical data for testing the relationships are not difficult to obtain. It is the purpose of the present paper to investigate some of the aspects of sediments which appear to follow exponential laws, to suggest simple methods of testing the applicability of the functions

[1] Some of these are mentioned in a later section of the paper.

to sets of data, and to consider some of the implications involved in the use of such functions.

In the more exact sciences, such as physics and chemistry, experimental data are more precise in general than are most geological data. This is due in large part to the fact that in the exact sciences it is possible to develop closed systems in which only certain factors are allowed to operate. By thus reducing the number of variables, it is a simpler matter to obtain data free from unknown influences. In geology it is practically impossible to find closed systems; even such relatively simple laboratory studies as mechanical analyses are complicated by the factors of grain shape, grain density, and the like, so that the resulting data do not have the simple geometrical significance that would be possible in a study of homogeneous spheres. In similar manner field measurements in geology are usually approximate and depend in large part on the accidental exposures that one finds. However, even with the inevitable scattering of values that obtain in the field and to some extent in the laboratory, it should be possible to test the data mathematically at least to a first degree of approximation; as more refined methods are developed for securing data the mathematical analysis may become more rigorous.

Fortunately there are available some quantitative data in sediments which are adequate for mathematical analysis in terms of exponential functions. The writer has been able to secure data from his own studies and from the literature to illustrate the method and to form the basis for a tentative theoretical discussion. The data to be used in this paper include such diverse items as the size of pebbles on a beach, the thickness of a loess deposit, the dispersion of glacial boulders in a boulder train, and the surface slope of an alluvial fan. Despite the relative paucity of data in these cases, the results are striking enough to warrant placing them on record. Moreover, the prevalence of a single type of analytical expression which may be applied to the variables in this wide range of sedimentary data suggests that exponential laws may have common application to many sedimentary situations. Some of the theoretical implications involved in exponential functions are discussed in a later section of this paper.

DECREASE IN SIZE OF BEACH PEBBLES

During the past year the writer, in collaboration with J. Scott Griffith, analyzed the beach deposits of Little Sister Bay, Wisconsin.[2] Although complexities were found among the deposits, due apparently to the influence of ice shove in the bay, the western half of the beach afforded data which may be used to illustrate the application of exponential functions in actual practice. The discussion will include methods of testing the applicability of the law as well as methods of determining the necessary constants.

Along the undisturbed portion of the beach at Little Sister Bay the average pebble size decreased as shown in Table 1. The distance,

TABLE 1

LITTLE SISTER BAY BEACH PEBBLES

Distance along Beach (feet)	Geometric Mean Diameter in Mm.	Distance along Beach (feet)	Geometric Mean Diameter in Mm.
0	52	240	27
70	43	290	25.5
140	35	350	21.5
190	31.5	555	13.5

expressed in feet from the westernmost sample, is indicated in the first column. The average size, shown in the second column, is the millimeter-equivalent of the first phi-moment,[3] and is physically the geometric mean diameter of the pebbles.

In this set of observational data we are interested in the change of average size as a function of the distance from any given point along the beach. Hence distance is the independent variable, x, and average size is the dependent variable, y. If the relation is exponential the average size should increase or decrease geometrically as the distance increases arithmetically. Figure 1 shows the curve obtained when average size is plotted against distance. The fact that the curve drops rapidly at first and then more slowly suggests a geo-

[2] "Beach Environment in Little Sister Bay, Wisconsin," *Geol. Soc. Amer. Bull.* (in press).

[3] W. C. Krumbein, "Application of Logarithmic Moments to Size Frequency Distribution of Sediments," *Jour. Sed. Petrol.*, Vol. VI (1936), pp. 35–47.

metrical relation, but it will be advantageous to test this more rigorously than by eye.

There are several methods of testing whether y changes geometrically as x changes arithmetically, and one of the simplest is by plotting log y against x. If such a graph is a straight line the relation holds.[4] Figure 2 is a graph of log y against x, made simply by plot-

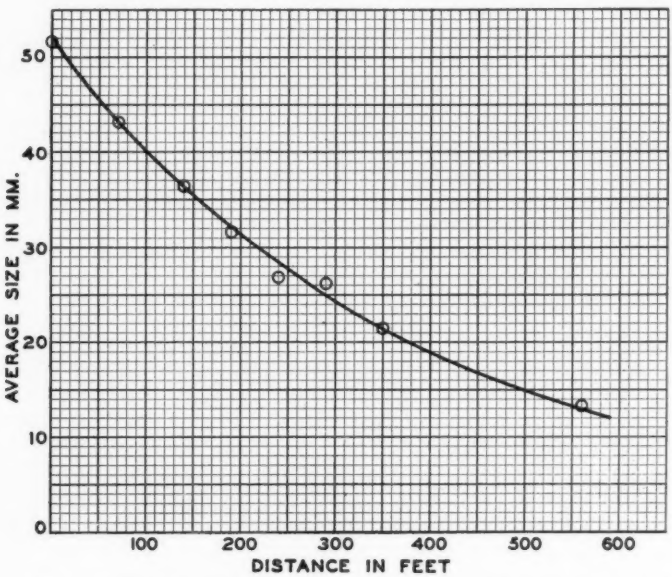

FIG. 1.—Graph of change in average size of pebbles along beach at Little Sister Bay, Wisconsin. Data from Table 1.

ting the data of Table 1 on semilogarithmic paper. The straight line drawn through the points indicates that the relation holds quite satisfactorily.

Inasmuch as the test for the exponential function appears to hold, and since y decreases as x increases, the relation between the variables is called a *negative exponential function*. Such a relation may be expressed by the equation $y = mb^{-ax}$, where m, b, and a are con-

[4] This follows from the fact that, when a variable changes geometrically, the logs of the values to any base change arithmetically.

stants. The constant b is called the *base* of the exponential function, m is the value of the dependent variable at $x = 0$, and is a constant, and a is called the *coefficient* of x, to which various names are given, depending on the physical situation involved. The negative sign before the exponent indicates that y decreases as x increases. Instead of using a purely arbitrary equation to express the relation between

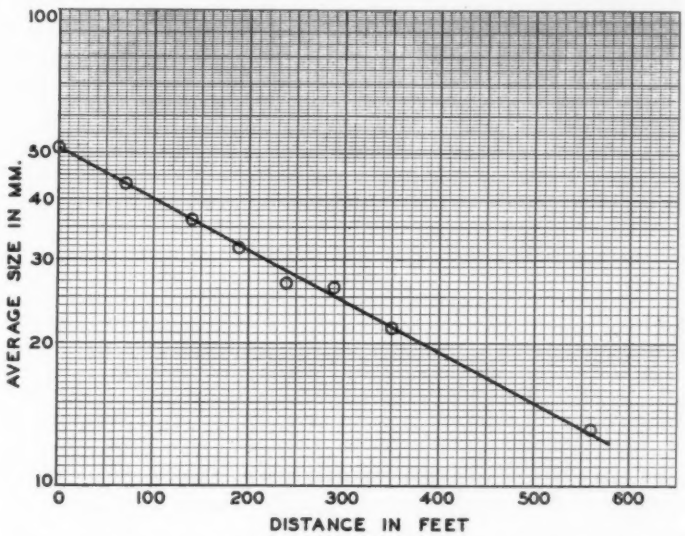

FIG. 2.—Semilogarithmic graph of change in average pebble size at Little Sister Bay.

x and y, it is more common to express the negative exponential as an equation in which for the base is chosen the number 2.7182, called e for convenience.[5] The constant m, from the nature of the equation, is chosen as the value of y at $x = 0$, which in the present case is 52 mm. This value is called y_0, or the value of y at the origin of the x-scale. In any given case y_0 is constant.

The reason for adopting these conventions is that, in developing the exponential function from a differential equation, the relations

[5] The number e is the base of natural logarithms and finds wide use in mathematical analysis.

are expressed directly in this form, as will be developed in the summary. When these conventions are followed the negative exponential equation becomes

$$y = y_0 e^{-ax}, \tag{1}$$

and the practical problem is reduced simply to the determination of the constant a.

There are several methods for determining a; one of the most convenient is a simple analytical method in which it is only necessary to find the value of x at the point where y is reduced to half its original value. In that case $y/y_0 = \frac{1}{2}$, and equation (1) becomes $\frac{1}{2} = e^{-ax_{1/2}}$. By taking logs of this expression to the base e, and changing the sign, there results $-\log_e(\frac{1}{2}) = ax_{1/2}$. But $-\log_e(\frac{1}{2}) = +\log_e 2 = 0.693$, and hence

$$a = \frac{0.693}{x_{1/2}} \tag{2}$$

where the value $x_{1/2}$ is called the "half-distance" and corresponds to the point where y is half its original value. In the present case y_0 is 52 mm., and the x-value at the point where $y = 26$ mm. (half its original value) is 260 feet, as determined from Figure 2. It will be convenient to choose 100 feet as the unit of distance, in which case the half-distance value of x is 2.6. Placing this value in equation (2) we obtain $a = 0.693/2.6 = 0.26$. Hence the required equation for pebble size is

$$y = 52e^{-0.26x}. \tag{3}$$

Here we may refer to the constant a as the *coefficient of pebble-size*. Physically this coefficient expresses the decrease in average size per mm. of diameter of the pebbles as they move unit distance (100 feet) along the beach. Geologically it may mean that the larger the pebble, the more rapidly it is reduced in size, but this point will receive further consideration below.

VARIATION IN THICKNESS OF LOESS

During the field season of 1936 a number of approximate measurements were made of the thickness of the Peorian loess in western

Illinois, in connection with another study.[6] The area extends southeastward from the Mississippi River, near the towns of Lomax and Dallas City, Illinois. It has long been known that the thickness of the loess decreases rapidly away from the bluffs along the Mississippi River and reaches a minimum thickness of about 5 or 6 feet over the greater part of the upland area. Unfortunately only a relatively few observations could be made in the thicker portions of the loess blanket, owing to a variety of complicating circumstances. Among these were the presence of dunes on the upland along the river bluffs and the absence of cuts deep enough to expose the entire section of

TABLE 2

THICKNESS OF PEORIAN LOESS

Distance from Bluffs in Miles	Thickness in Feet	Distance from Bluffs in Miles	Thickness in Feet
2.0	19±	7.5	7+
2.9	14±	7.8	8+
3.8	12	8.0	6
4.6	10+	9.0	5
4.8	12±	9.1	6
5.2	11	10.0	7±
5.4	10+	10.7	6
6.0	10+	11.4	7
6.1	9±	13.2	6

loess at points close to the river. In addition the region close to the Mississippi River is considerably dissected, and it is difficult to judge whether one has the entire deposit in a given exposure. Despite these shortcomings, the data available are perhaps sufficient to test whether an exponential function is indicated. Table 2 is a compilation of loess thicknesses observed at varying distances from the Mississippi River bluffs. The observations do not lie in a single straight line but are scattered over the area studied.

In this example two problems are involved. First, whether the decrease in thickness is exponential and, second, how thick the loess itself is along the river bluffs. Figure 3 is a graph of the data in Table 2 plotted on semilogarithmic paper. A straight line has been

[6] A study of areal variations in the characteristics of loess, supported by a grant from the Illinois State Geological Survey, Dr. M. M. Leighton, chief.

drawn through the points. It will be noticed that in this case the points scatter more widely than in the case of the beach pebbles, but in general there is a suggestion that an exponential function is involved. In most cases of geological field data, where observations are made on random exposures, one may expect a scattering of values. In the present observations there are such factors as possible

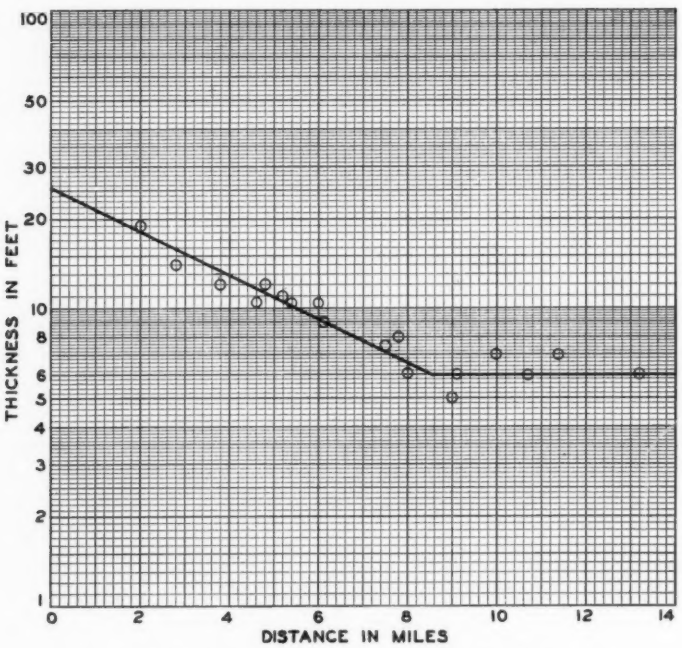

FIG. 3.—Semilogarithmic graph of loess thickness. Data from Table 2

erosion since deposition, irregularities of the underlying till plain, and possible minor contributions from smaller flood plains in the area.

It will be noticed that beyond the point $x = 9$ miles the curve approximates a horizontal straight line, indicating that the exponential function no longer holds, but that a more or less uniform blanket of loess lies over the upland. This feature is common to loess deposits

over much of the upland area studied, and probably depends on the nature of dust transportation by wind. It would seem that close to the source the loess settles out partly on the basis of relative grain size, but that, beyond the point where significant size differences are present, the fine dust settles more or less uniformly. If this relation holds, then the average size of the loess grains should decrease appreciably over the first 9 or so miles from the river bluffs. It may even be expected on theoretical grounds that the size decrease is also exponential. No data are available on this question, however.

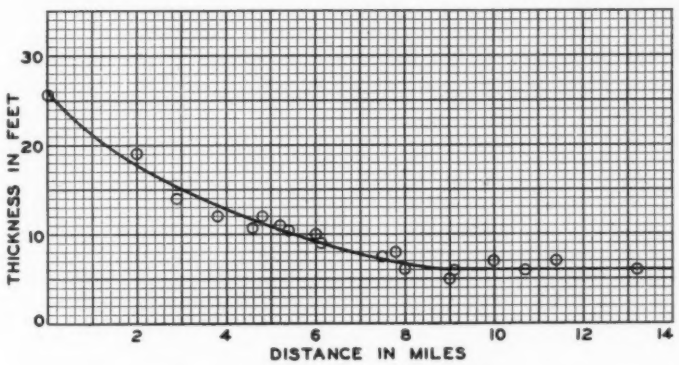

FIG. 4.—Comparison of observed and theoretical thickness of loess

To determine the thickness of the loess along the river bluffs, the straight line of Figure 3 may be projected to the y-axis. In the figure the line intersects the axis between 25 and 26 feet. Using the value 26 as y_0, Figure 4 was drawn, which shows the exponential nature of the curve directly.

The thickness of the bluff loess as determined by the graph seems low in comparison with the observable height of the bluffs themselves. However, when the situation along the bluffs is analyzed, it is found that sand dunes occur as a fringe along the tops of the bluffs, and road cuts along the bluff slopes show numerous layers of sand interbedded with the loess. Thus the bluffs themselves are undoubtedly an accumulation of wind-blown sand as well as loess. This situation is to be expected near the source of wind deposits when it is re-

called that wind transportation often involves a traction load of sand as well as a suspension load of dust. The traction load is stopped relatively soon by surface obstacles, whereas much of the suspension load continues on. Thus at the bluffs themselves the total accumulation of deposits may be much more than 26 feet; indeed, at various localities along the bluffs it has been estimated that the loess is about 40 feet thick. On the basis of the current study about one-third of this would be sand and two-thirds true loess.

The determination of the constant a may again be accomplished by the analytical method, using equation (2). The half-distance may be read from Figure 4, which indicates that a thickness of 13 feet (half of y_0, 26 feet) occurs about 4 miles from the bluffs. Hence $a = 0.693/4 = 0.17$. The equation for the decrease in loess thickness is accordingly

$$y = 26e^{-0.17x}. \tag{4}$$

The constant a may be called the *coefficient of loess deposition* or, more generally, the *coefficient of loess thickness*. Physically the value of the coefficient indicates that the thickness of the loess decreases by a decrement of 17 per cent of itself per mile.

PROFILE OF AN ALLUVIAL FAN

Among the topographic maps which display alluvial fans, the Cucamonga, California, sheet has long been a classic. A portion of this map is shown in Figure 5, with a line drawn from the mouth of San Antonio Canyon southward down the alluvial fan. The elevation of the surface of the fan was read at mile intervals along this line to a point beyond the edge of this figure, as shown on the entire map itself. The data are shown in Table 3, and Figure 6 is a semilogarithmic graph of the data. Figure 7 shows the surface profile itself with the vertical scale exaggerated to bring out the curvature.

It will be noted that the logarithmic graph yields a rather satisfactory straight line, and from Figure 7 the half-distance is found to be approximately 5.6 miles.[7] Using equation (2), the constant a is

[7] As a first approximation the half-distance is taken where $y = 1{,}140$. If the half-distance is based on half the difference between the highest and lowest elevations (2,280 and 770), the value of a will be larger.

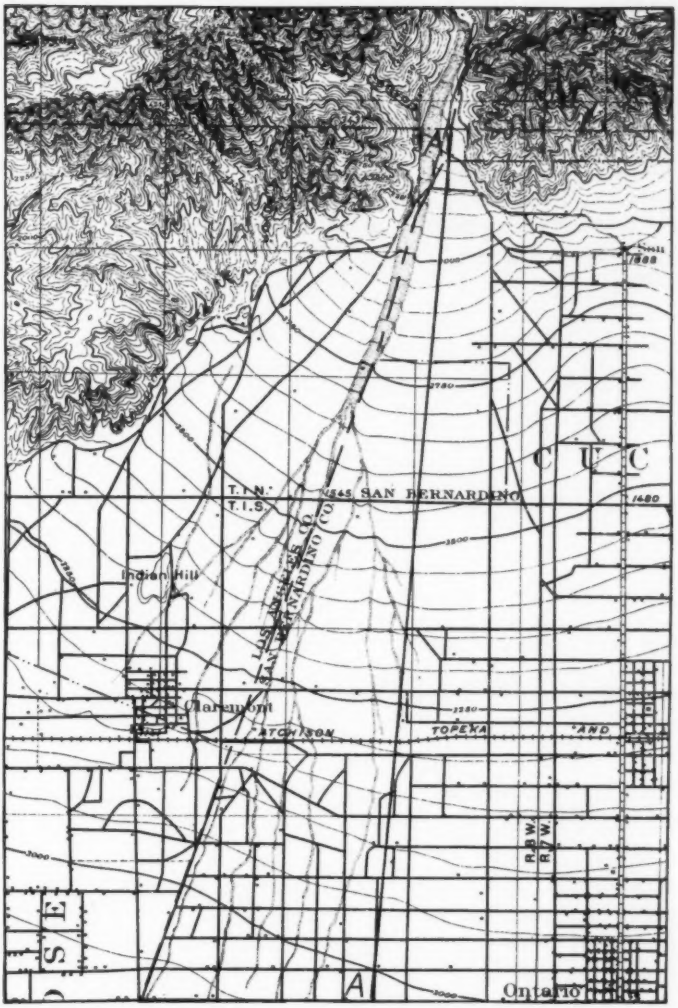

FIG. 5.—Portion of the Cucamonga, California, topographic sheet. The line *AA* indicates profile measured.

found to have the value $a = 0.693/5.6 = 0.12$. The equation of the profile is therefore

$$y = 2{,}280e^{-0.12x}. \tag{5}$$

The constant a may be called the *profile coefficient*, but it is also a coefficient of sediment thickness because, with reference to the surface of deposition, the profile of the alluvial fan is an expression of the thickness of the deposited material at any point. Hence the constant a is also a coefficient of deposition. Physically the value of the coefficient signifies that in this alluvial fan the surface elevation de-

TABLE 3

SURFACE ELEVATION OF SAN ANTONIO ALLUVIAL FAN

Distance from Mouth in Miles	Elevation in Feet A.T.	Distance from Mouth in Miles	Elevation in Feet A.T.
0	2,280	6	1,090
1	2,000	7	990
2	1,760	8	900
3	1,560	9	835
4	1,370	10	770
5	1,210		

creases by a decrement of 12 per cent of itself per mile, with reference to the original surface of deposition; or, from a thickness point of view, that the thickness of the alluvial deposit decreases by a decrement of 12 per cent of itself in each mile.

It may be well to point out explicitly that the mere circumstance that a single alluvial fan appears to follow an exponential law is no demonstration that all of them do. The same argument obviously applies to the size of beach pebbles and the thickness of loess. It will also apply to the case of boulder concentration in a glacial boulder train, to be introduced shortly. Nevertheless it will be shown in the theoretical discussion that it is not improbable that exponential curves may well be a commonplace in sedimentary situations. In fact, it should also be pointed out that even if alluvial fan data, for example, do not all plot as straight lines on semilogarithmic paper, it may not be safe to assume that exponential functions do not apply. If the alluvial fan is deposited on a slope, the surface of the fan

may itself be exponential when referred to the angle of slope of the original surface, but when the data are plotted logarithmically with respect to a horizontal axis, the curve is not a straight line. In like

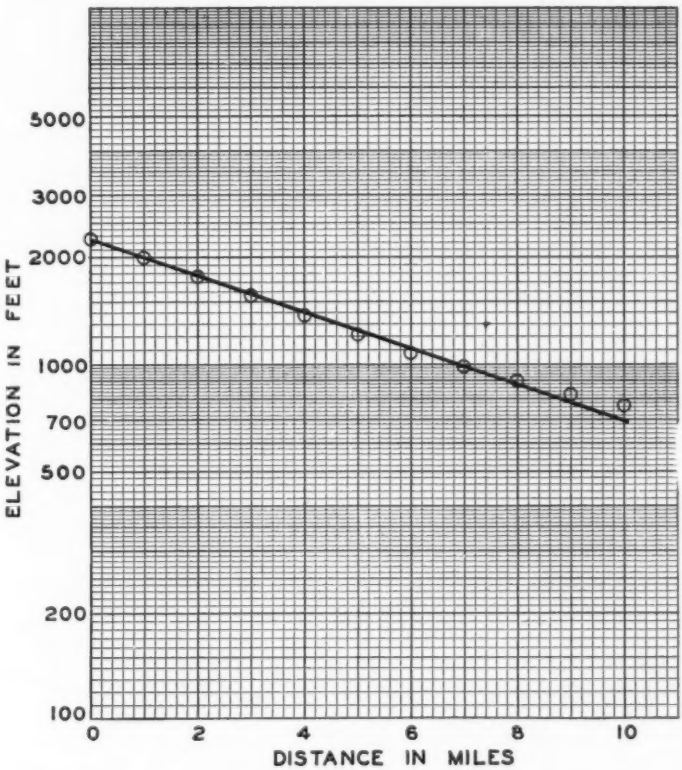

FIG. 6.—Semilogarithmic graph of surface elevation of San Antonio Canyon alluvial fan. Data from Table 3.

manner complexities may arise in connection with tectonic disturbances which may result in the superposition of one fan above the other. Fortunately such complexities appear to be amenable to attack by graphic methods.

DISPERSION OF GLACIAL BOULDERS IN A BOULDER TRAIN

It has long been known that the boulders along a glacial boulder train thin out as one moves away from the source outcrop. The question arises whether such a decrease in the concentration of boulders per unit area may also be an exponential function. A search of the

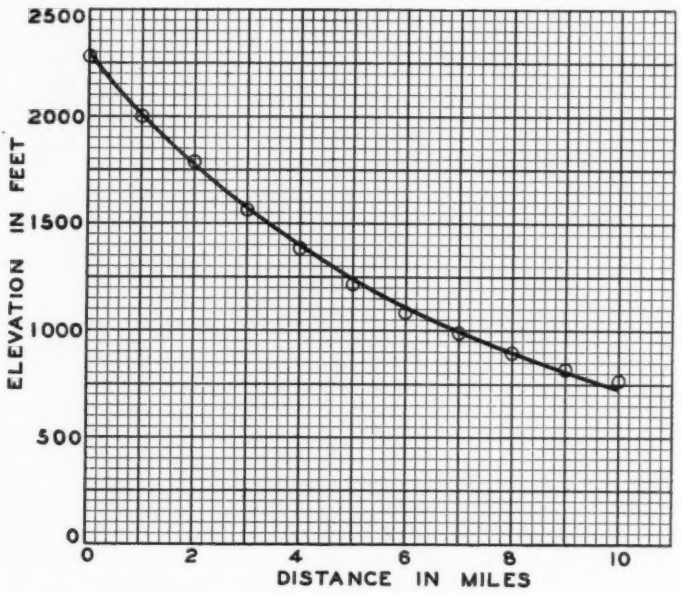

FIG. 7.—Comparison of observed and theoretical surface elevation of San Antonio Canyon alluvial fan.

literature disclosed a map of the Mount Ascutney boulder train of New Hampshire and Vermont[8] which showed the location of individual boulders. Figure 8 is a sketch based on the original map, to illustrate the situation. The boulder train was divided into concentric segments 10 miles wide, and a count of the boulders in each segment is shown in Table 4. To determine the number of boulders per unit area, the innermost sector was considered to have unit area, and

[8] J. W. Goldthwait, "The Geology of New Hampshire," *N.H. Acad. Sci. Hdbk. 1* (1925), map by F. H. Foster, p. 20.

the successive outer segments then have areas 3, 5, 7, and 9 times that of the inner sector.[9] By dividing the frequencies in the second column of Table 4 by these successive numbers, the concentration of boulders per unit area was found, as shown in the third column of the table. Figure 9 shows these values plotted on semilogarithmic paper with the straight line *A* drawn through them.

At first glance the line *A* appears to fit satisfactorily, but several complexities arise in choosing any single straight line through the observed points. This is due to the relative paucity of data in the outer zones of the boulder train. Six observations are hardly enough to afford a critical value. Statistically, if a number of similar boulder trains were mapped, it may be expected that the observed values in the outer zones would show rather wide fluctuations. Hence it is not safe to assume, even if the actual number of boulders is six in the given case, that this number has a high significance. Indeed, it is likely that more weight should be given to the observed values in the inner zones because of the greater number of boulders present. On this basis the line of best fit is the lower light line in Figure 9. On the other hand, if by a small probability the boulders found in the outer zones should be the expected numbers, then the upper light line in the figure may be the better fit.

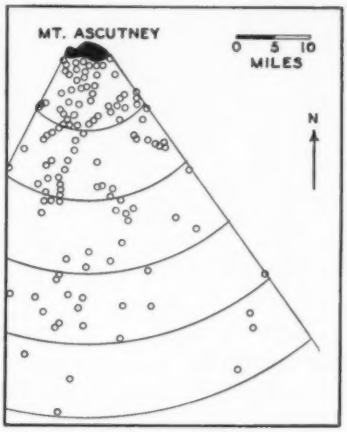

FIG. 8.—Sketch map of Mount Ascutney boulder train, after Foster. Each circle represents a boulder; the arcs are spaced at 10-mile intervals.

The point to this discussion is that, in the event of meager data

[9] This follows from the fact that, when the area of a sector of a circle is divided into segments of equal width, the areas of the segments successively outward follow the odd integer law. This may be proved by recalling that the areas of circles (or sectors) vary as the square of the radii, so that the area of a sector of radius 1 is one-fourth that of a sector of radius 2. Hence the difference of areas is 1: 3, and so on for radii 3, 4, 5. . . . .

available for a given geological phenomenon, it may not be safe to assume that the line of apparent best fit is theoretically the most significant or that any single straight line satisfies the conditions imposed by the limited data. This, it should be borne in mind, is independent of the fact that the observer has found every boulder in the area. The question is not one of accuracy of field work; it is simply a problem of error magnitude imposed by the data themselves.

Despite the uncertainties which attach to the data, it appears to the writer that a straight line is a better fit to data of this nature than

TABLE 4

DISTRIBUTION OF BOULDERS IN MOUNT ASCUTNEY BOULDER TRAIN

Zone	Total Boulders	Relative Area of Zone	Boulders per Unit Area
1	47	1	47.0
2	35	3	11.6
3	17	5	3.4
4	16	7	2.3
5	6	9	0.67

any curved line. Hence it appears reasonable to assume that an exponential function may apply, but in this case no single value of the constant a can be chosen as satisfying the data uniquely. However, if the two extreme straight lines are considered as limiting a zone in which the correct function lies, it is possible to express the range of values which the constant may be expected to have in this case.

From the upper line in Figure 9 the half-distance is 7 miles, so that the constant a is found from equation (2) to be $a = 0.693/7 = 0.10$; similarly, from the lower line in the figure the half-distance is 5 miles, which yields the value $a = 0.693/5 = 0.14$. (The line A itself yields the value $a = 0.11$.) From the range of values found it may be taken as a first approximation that the function is

$$y = 47e^{-(0.12\pm0.02)x}. \tag{6}$$

Figure 10 shows the curves from Figure 9 plotted as negative exponentials directly, to show the range of possible curves. It should be noted that it is immaterial what value is assigned to the origin;

the boulder concentration, y_o, drops to half its original value in the range from 5 to 7 miles from the chosen origin.

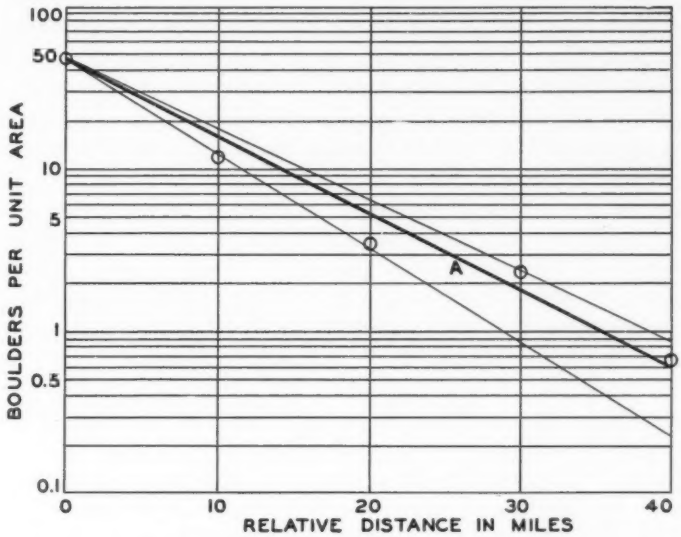

FIG. 9.—Semilogarithmic graph of boulder concentration per unit area in Mount Ascutney boulder train. Data from Table 4.

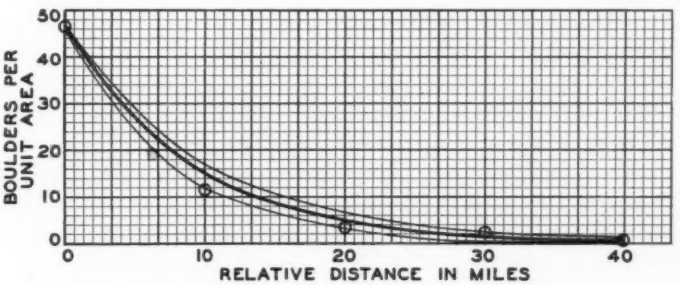

FIG. 10.—Comparison of observed and theoretical boulder concentrations per unit area in Mount Ascutney boulder train.

On the assumption that the function involved is a negative exponential, the physical significance of the constant a is that the

boulder concentration drops by a decrement of about 12 per cent of its value every mile. The constant may therefore be called the *coefficient of boulder concentration*. There is another manner of viewing the constant in this case, however. If the boulders per unit area were piled one above the other, they would form columns of varying height, which would be analogous to the thickness of the boulder deposit over the area. In this sense the coefficient could be considered as a coefficient of thickness, as in the cases of the loess and the alluvial fan.

In the last three cases, then, there is a physical similarity between the various coefficients. It is to be noted that the deposits were respectively laid by wind, running water, and glacial ice. Whether there is any necessary relation between the underlying laws of deposition by these three widely different geological agents will be considered in the following section.

THEORETICAL DISCUSSION

The several geological situations discussed above offer material for a tentative discussion, but it will be necessary to examine many more situations before valid generalizations may be made. The writer believes that the several examples cited offer suggestions of the wide field available for study in connection with exponential laws. At present sedimentary studies are tending more toward areal or linear sets of samples, and the data available from such studies should afford additional critical evidence of the nature of the underlying functions.

One may find in the literature here and there actual applications of exponential functions to sedimentary data. For example, in 1875 Sternberg[10] developed his abrasion law in which it was shown that the weight of a stream pebble is proportional to the work done in overcoming friction over the distance traveled. In 1925 Barrell[11] expressed this law directly as a negative exponential of the type $W/W_0 = e^{-\lambda l}$, where W, λ, and l are respectively the weight of the

[10] H. Sternberg, "Untersuchungen über Längen- und Querprofil geschiebeführende Flüsse," *Zeit. f. Bauwesen*, Vol. XXV (1875), pp. 483–506.

[11] J. Barrell, "Marine and Terrestrial Conglomerates," *Geol. Soc. Amer. Bull. 36* (1925), p. 328.

pebble, the coefficient of abrasion, and the distance. More recently Shulits[12] used Sternberg's law to develop a negative exponential for expressing the equation of river slope. Wentworth[13] performed experiments on the abrasion of limestone cubes and showed that his results plotted as straight lines on semilogarithmic paper. In 1930 Athy[14] demonstrated that the porosity of a shale varied exponentially with depth of burial in accordance with the equation $P = pe^{-bx}$, where P, b, and x refer, respectively, to the porosity, the coefficient, and the depth of burial. In connection with the relation between turbulence and the distribution of suspended matter in equilibrium in a column of fluid, Leighly[15] cited Schmidt's law, which is a negative exponential. It is also interesting to note, in connection with the glacial boulder train, that as long ago as 1900 Salisbury[16] discussed the local origin of glacial drift, and in setting up theoretical values for the dispersal of the drift he arrived at a qualitative statement which may be approximated by a negative exponential.

These several examples indicate that there is a wide range of sedimentary phenomena in which negative exponentials are involved, and unquestionably many other situations will be found or have already been found. Not alone is this true of sediments, but very likely many general geological phenomena show exponential functional variations, either with respect to time or with respect to distance.

In order to seek for the underlying factors which may control exponential functions, it will be well to consider some of the properties of exponential curves in general. In any exponential function the rate of change of the dependent variable is always proportional to the value of the dependent variable at the point under consideration.

[12] S. Shulits, "Fluvial Morphology in Terms of Slope, Abrasion, and Bed-Load," *Trans. Amer. Geophys. Union*, Part II (1936), pp. 440–44.

[13] C. K. Wentworth, "Pebble Wear on the Jarvis Island Beach," *Wash. Univ. Studies, New Series, Sci. and Techn.*, No. 5 (1931), pp. 11–37.

[14] L. F. Athy, "Compaction and Oil Migration," *Bull. Amer. Assoc. Petrol. Geol.*, Vol. XIV (1930), pp. 1–24.

[15] J. Leighly, "Turbulence and the Transportation of Rock Debris by Streams," *Geog. Rev.*, Vol. XXIV (1934), pp. 453–64.

[16] R. D. Salisbury, "The Local Origin of Glacial Drift," *Jour. Geol.*, Vol. VIII (1900) pp. 426–32.

In order to develop a picture of the physical meaning of an exponential function, and especially of the physical significance of the constant a, it is necessary to consider the general mathematical procedures by which exponential functions are developed. Usually a differential equation is set up which enables us to interpret the physical meaning of each term in the equation. The fact that the rate of change of the dependent variable is proportional to the value of the ordinate at any point is expressed as

$$dy = -aydx \,, \tag{7}$$

where dy is the change in the dependent variable, dx is the change in the independent variable, y is the value of the ordinate (dependent variable) at any point, and a is the constant of proportionality. The negative sign indicates that y decreases as x increases, and thus defines the negative exponential.

The differential equation (7) is readily solved by rearranging the symbols into the form $dy/y = -adx$. By integration this yields $y = -ax + \log C$, where $\log C$ is the constant of integration. To evaluate this constant, let $x = 0$, whereupon $\log y = \log C$; this demonstrates that C is y_0, the value of y at the origin. Hence the equation becomes $\log y = -ax + \log y_0$, or $\log (y/y_0) = -ax$. By taking the antilog of the last expression we obtain

$$\frac{y}{y_0} = e^{-ax} \,, \tag{8}$$

which is the equation used throughout the text when y_0 is carried to the other side of the equation. The constant a is thus really a constant of proportionality, and in its position in the final equation it is called the coefficient of the independent variable. In choosing a name for the coefficient in any case, the writer believes that a general term is preferable to any specific name which relates it to particular processes, as was stated in connection with the beach pebbles.

In any event the value of a must be constant in order for the exponential law to hold, and this requirement raises interesting ques-

tions. For example, when one considers the many variables involved on a beach, such as the strength of the waves, the undertow, the slope of the beach, the density of the pebbles, their degree of sphericity and degree of roundness, and perhaps others, it is surprising that the net effect of these is a constant which expresses quantitatively the change in size. When one considers the many factors in different kinds of environments, and their variation from place to place and from time to time, it is almost unbelievable that so simple a law as the negative exponential could hold.

In addition to the constant a there are at least two other values associated with exponential curves which have geometrical significance. These are the "half-distance value" and the "mean distance value" of the function. The half-distance, $x_{1/2}$, has already been defined as that distance at which $y/y_0 = \frac{1}{2}$; in other words, where the thickness (or size) is half its original value. The mean distance, x_m, is the x-value of the centroid of the curve, and it may be shown that $x_m = 1/a$. The mean distance is thus an arithmetic mean value (the value of the first moment of the function), and numerically it is equal to the reciprocal of the constant a. Finally, the relation between the half-distance and the mean distance is $x_{1/2} = x_m \log_e 2$. The particular physical interpretations which may be placed on these values differ according to the nature of the data and depend upon the theory chosen to explain a given set of observations. The point will receive consideration below. However, one interpretation of x_m may be given to illustrate the point. When one considers merely the thickness of a deposit, such as the loess, it is possible to use x_m to compute the total volume of material deposited over the area. The area under the exponential curve is given by the equation $A = y_0 x_m$, and hence the volume in a given strip may be found by multiplying the area by the width of the strip. Thus, for the loess, $y_0 =$ 26 feet, $1/a = 1/0.17 = 5.87$ miles = 31,000 feet, and the area under the curve is 807,000 square feet. In a strip one foot wide there will accordingly be 807,000 cubic feet of loess, on the assumption that the exponential law holds throughout.

There are perhaps two broad situations which may be considered in connection with exponential functions. In one case there is a single controlling factor of the environment which determines the

law and the value of the coefficient a. In the other case there are numerous independent factors operative, none of which is predominant. Curiously enough, this latter situation sometimes gives rise to an exponential function. In this case, however, the constant a cannot be attributed to any single factor, but its value must be considered to be due to the interplay of all the factors. It is not feasible to say with certainty which of these two situations holds in any given case, but some speculations may be permitted.

In the case of the beach pebbles it is not unlikely that the value of the constant a depends upon more than a single factor. Wave action is probably the principal agency which moves the pebbles to and fro along the slope of the beach. The energy of the waves themselves varies, however, so that in some cases the pebbles are rolled along and in others they may literally be tossed. The first situation probably gives rise to abrasion; the second may result in chipping and breaking. The net effect is size reduction, and this size reduction appears to depend partly on the ease with which the pebbles are broken or worn. Bedding planes may play a part here. The fact that the decrease in size is rapid at first and less rapid later on indicates that the larger the pebble, the more rapidly it is reduced in size.

On the other hand, one must not lose sight of an alternative theory which involves selective transportation instead of size reduction. Thus abrasion, chipping, etc., imply an agent which changes the size of individual pebbles. However, some of the pebbles may retain their given size, but merely the smaller ones may be transported farther (on the average) than the larger. There seems to be, then, a choice between an actual decrease in average size due to breakage or abrasion or an apparent decrease in size due to selective transportation. The truth probably lies between. The following several alternatives appear to be involved: In terms of sediment thickness there is a simple transportation theory which involves particles all of one size, with no abrasion whatever, and a selective transportation theory which involves a distribution of sizes and a selective effect during transport, but again no abrasion. A third possibility involves abrasion alone, with no selective transportation effects; and, finally, there may be a combination of abrasion and selective transportation. In each of these cases there will be restrictions on the

physical meaning of such values as the constant a, the mean distance, x_m, the half-distance, $x_{1/2}$, and other values. The restrictions will be imposed by the assumptions on which the defining differential equations are set up.

The lack of sufficient experimental evidence regarding the several alternative theories suggests that the functional relationship of size to distance should be treated as a broad generalization, rather than reading into the observations a theory which may or may not be valid. One may hesitate therefore, to call the beach pebble law an "abrasion law" or a "selective transportation law" until more data are at hand. Rather, by generalizing the relations as a law of the change in average pebble size along a beach, the analytical expression would still make quantitative the fact that beach pebbles decrease in average size in accordance with a negative exponential function.[17] In similar manner, when the varying thickness of a sedimentary deposit is considered, the exponential law may be expressed in general terms to avoid using any specific geological process as a postulate of the law. Thus, although size, shape, and density of grain may be involved, as well as the nature of flow of the medium (turbulent or laminar), the law that some deposits may decrease in thickness exponentially may be expressed descriptively as a law of thickness variation.

The foregoing discussion demonstrates that, in addition to a choice of underlying theory, there may be a wide variety of factors involved in the value assumed by the constant a. Thus, although negative exponentials may be fairly simple mathematically, a given function as applied to sediments may be quite complex. Fortunately methods are available for resolving some of these complexities. Even though several independent variables may be combined in the value assumed by the constant a, it is possible that some dominating factor (wave action, for example) may control the order of magnitude of the constant in a given case. In such cases the exponential function may sometimes be expressed as $y = y_o e^{-(a_1+a_2)x}$, where $(a_1 + a_2) = a$, and

[17] It should be borne in mind that the beach-pebble size-change law is a statistical law. That is, the observations disclose that the geometric mean size of the pebbles decreases exponentially with distance. In a strict sense, then, the observations do not relate to individual pebbles, which may depart from the law of the average.

a_1 is the contribution of wave action, whereas a_2 is due to other factors. The situation may be generalized to $a = a_1 + a_2 + \cdots + a_n$, for as many factors as may be experimentally isolated and determined.

THE ROLE OF PROBABILITY IN EXPONENTIAL FUNCTIONS

Exponential laws find wide application in the study of radioactive decomposition, and one may refer to Rutherford's treatise[18] for an approach to exponential theory based on probabilities. Rutherford's fundamental law of radioactive decay does not concern itself with the assignment of specific causes for the decay or explosion of a given atom. It merely asserts (1) that it is equally probable that a given atom of radium, say, will decay in the next second or during a given second 100 years from now, and (2) that this probability of decay, λ, is the same for all radium atoms. From these two specific laws the relation $N = N_0 e^{-\lambda t}$ is derived, where N_0 is the number of atoms of radium in a given sample at time $t = 0$ and N is the number left after t seconds.

This derivation, it will be noted, defines λ as the probability that an atom shall be transformed in unit time. If the same reasoning is applied to sedimentary situations, one may consider the constant a as representing the probability that a given pebble will be reduced in size per unit distance, or that a given pebble will be deposited in unit distance, depending upon the point of view applied. The sedimentary situation by this probability assumption is converted into a statistical problem, amenable to attack by statistical methods. Furthermore, chance deviations from the theoretical may then be evaluated and some notion obtained of the operation of the laws of probability in sedimentary situations. *On such a statistical basis one could view any given sample of sediment as the most probable deposit at that time and that place, in terms of the environmental factors operating and the range of materials present.*

The implications of this line of reasoning may readily be seen in connection with the analysis of ancient deposits. If the value of the constant a is controlled in part by probability considerations, it may

[18] Sir Ernest Rutherford, James Chadwick, and C. D. Ellis, *Radiations from Radioactive Substances* (Cambridge, 1930), pp. 167 ff.

be difficult or impossible to evaluate the contributions of any single factor in the environment. With present-day deposits, on the other hand, the use of the equation $a = a_1 + a_2 + \cdots + a_n$ may permit a detailed analysis of the effect of any factor which may be experimentally isolated. It is also possible that these results may be carried over into the study of ancient deposits, although in some cases it may be anticipated that a residuum of uncertainty may remain in the complete evaluation of ancient sedimentary environments.

Aside from the possible complexities introduced into the study of ancient sediments by chance deviations about a given expected value, the very fact that the coefficient a is a constant may in some cases permit a simple analytical expression for geological phenomena, despite the underlying complexity. For example, if it could be established that certain characteristics of given sedimentary deposits, without regard to agent, vary exponentially as a function of distance, that generalization would in itself be important, even though the factors that control the situation in any given case may not be known.

The principal function of the present paper is to present a line of reasoning suggested by earlier work on exponential functions and by additional data on sedimentary deposits. The fact that these several phenomena suggest exponential functions, at least to a first degree of approximation, affords opportunity for a more quantitative examination of geological data and points a possible way for simplifying the underlying relations among a wide variety of phenomena. The writer would like to have the paper accepted as an approach along this avenue of thought. It is also a pleasure to acknowledge the aid received from Dr. Carl Eckart of the Department of Physics of the University of Chicago regarding the physical significance of exponential functions.

REVISED CLASSIFICATION OF MARINE SHORELINES

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ABSTRACT

In former classifications of shorelines emergence and submergence have been used as the most important elements. However, the shifting sea levels of Quaternary time with the present intermediate condition could have left indications of both emergence and submergence along almost any coast. Furthermore, the criteria commonly used to classify coasts are in part open to question. A new system is introduced in this article which attempts to overcome many difficulties of the older classifications. While this system is entirely genetic, it was devised so as to allow its ready application to many coasts merely from the information which is contained in maps and charts.

INTRODUCTION

Many of the early attempts to classify shorelines were little more than descriptions of coasts. However, with the work of Gulliver[1] and Davis[2] these older empirical classifications began to be replaced by genetic classifications. This desirable change was pushed further by Johnson[3] producing what has been a widely utilized classification for almost a score of years. This system of Johnson's includes shorelines of submergence (ria and fiord types), shorelines of emergence, neutral shorelines (delta, alluvial plains, outwash plains, volcanoes, coral reefs, and fault shorelines), and compound shorelines.

OBJECTIONS TO PREVIOUS CLASSIFICATIONS

Importance of glacial control.—The foregoing classification may seem on first thought to be quite ideal, and in view of the small amount of information available concerning the oceans it was no doubt logically conceived. Now, however, that we are beginning to uncover some of the history of the oceans it seems necessary to offer some revisions. Perhaps the principal conflict comes from the

[1] F. P. Gulliver, "Shoreline Topography," *Proc. Amer. Acad. Arts and Sci.*, Vol. XXXIV (1899), pp. 151–258.

[2] W. M. Davis, *Geographical Essays* (Boston, 1909), p. 178.

[3] D. W. Johnson, *Shore Processes and Shoreline Development* (New York, 1919), pp. 159–92.

establishment of the importance of glacial control of sea-level, an idea which has long been advocated by Daly and other scientists. According to this theory the sea is at present 100 feet or perhaps even 200 feet lower than it would be if all the glaciers should melt.[4] Therefore, since these glaciers have not always existed all coasts are potentially shorelines of emergence. On the other hand during the glacial period the great growth of the ice caps caused a lowering of the sea-level of hundreds of feet below the present and perhaps of thousands of feet.[5] From this second point it becomes evident that all shorelines are potentially shorelines of submergence. In other words, if a coast has had no diastrophic movements whatsoever it may show, on the one hand, wave-cut terraces at various levels up to about 200 feet above sea-level, and, on the other hand, it may be indented by estuaries or be bordered by submarine canyons of fluviatile origin. Such a coast might be interpreted as a coastline of emergence by an observer who was impressed by the terraces, while the estuaries or the submarine canyons might lead another observer to term the shoreline as one of submergence; and still another more thorough student could term the same coast as a compound shoreline in the sense that it had indications of having both emerged and submerged. Examination of charts and perusal of the literature will show that practically all coasts have these indications both of emergence and submergence. Surely a classification in which one member comprises practically all cases is inadequate.

Straight coasts and emergence.—In addition to the elevated wave-cut terraces, a criterion of emergence which one may find in most textbooks is straightness of coastline. A little reasoning will show the danger of making any such generalization. If we take a section of the California coast (almost any part would do, but Fig. 1 is representative) we may observe that the coastline is much straighter than any of the adjacent submarine contours. Therefore an uplift at the present time instead of straightening this coast would make

[4] For discussion of the amounts see R. A. Daly, *Changing World of the Ice Age* (New Haven, 1934), pp. 170–76.

[5] F. P. Shepard, "Underlying Causes of Submarine Canyons," *Proc. Natl. Acad. Sci.*, Vol. XXII (1936), pp. 496–502.

it more crooked. Furthermore, upon making a careful examination of the same coast it is evident that there are indications of former estuaries, but that coastal deposition combined with deposition of rivers has straightened the coast. In other words, its straight condition has been due to its relative stillstand in recent years rather than to any recent emergence.

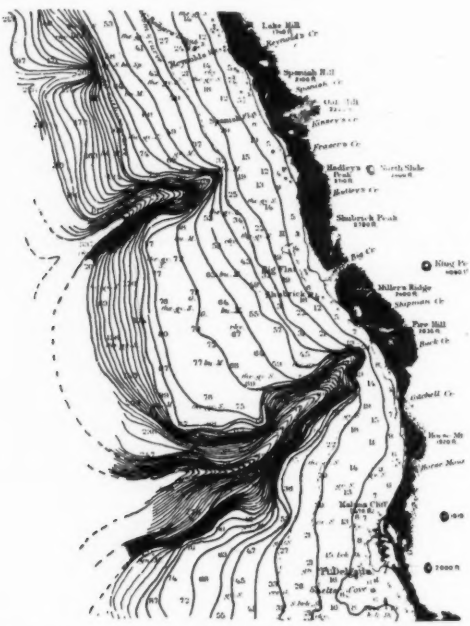

FIG. 1.—Section along the California coast near Cape Mendocino illustrating the effect on the coastline which would develop due to emergence. (Coast and Geodetic Survey Chart 5700.) Contoured by the late Professor H. H. Robinson.

The foregoing illustration is typical and could be duplicated from many other coasts of the world, although some localities would offer examples of decided straightening which might result at least in part from uplift.

Offshore bars and emergence.—On the plea that the continental shelves are relatively flat and that offshore bars form where there

is a very gentle seaward slope, Johnson has claimed that these bars are indicative of a coast line of emergence.[6] This criterion also needs careful scrutiny before it should be used. Assuming that offshore bars develop where there is a gentle seaward slope, would not such a condition be produced by the submergence either of a coastal plain or of a delta? Certainly the continental shelf is not the only flat surface. Furthermore are there not many examples of barrier beaches found off coasts which are deeply indented by estuaries? The coast of North Carolina is a clear example of such a combination (Fig. 2).

PROPOSED CLASSIFICATION

Having suggested difficulties with present classifications it remains to offer a substitute. This is no easy task and the writer realizes that any proposed system will be likely to have serious weaknesses. Nevertheless it is time that someone ventured a new classification with the hope that others will point out ways of improving it. In making the new proposals, first, we should include all desirable elements of the old classifications; second, we should set up a system which would take into account the world-wide sea-level changes during the glacial period; and, third, we should develop groups which are of easy application. The recency of the melting of the glaciers and the rise in sea-level should give us the basis for dividing the coasts into two major groupings, namely, (1) primary coasts, those where the waves have not had time to produce notable effects; and (2) secondary coasts, those where marine processes have been able to shape the coasts. Under the first grouping we will include coasts which owe their form principally to nonmarine agencies. Further subdivision under the first group will be made on the basis of the process which has been responsible for the land form against which the sea has come to rest. Division of the second group will be made by consideration of the marine process which was involved in shaping the coast. In such a classification there must of course be many gradational cases, both between the first and the second large groupings and between the various subdivisions of each. There is no reason, however, why two classes

[6] D. W. Johnson, *op. cit.*, pp. 348–92.

should not be used for the same coast, as, for example, a ria coast with offshore bars. Also an examination of various charts and a

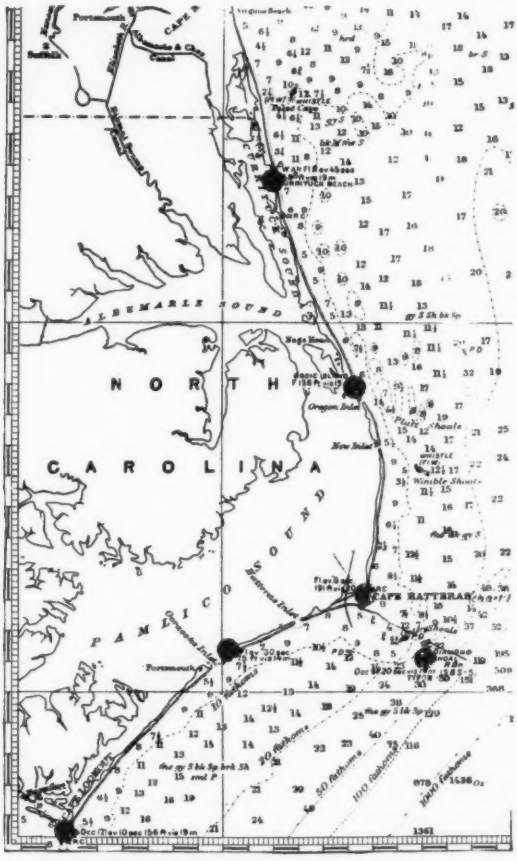

FIG. 2.—Illustrating the combination of barrier beaches and estuaries. (Coast and Geodetic Survey Chart 1000.)

study of many coasts in the field has shown that the list which will be given below can be applied easily to most coasts.

The classification (Table 1) is in part the result of an attempt

TABLE 1—*Continued*

B. Coasts shaped by marine deposition
   1. Coasts straightened by building of bars across estuaries (Fig. 20)
   2. Coasts prograded by wave and current deposits (principally in bays) (Fig. 21)
   3. Coasts with offshore bars and long shore spits (Figs. 22 and 23)
   4. Coral reef coasts
      a) Fringing reefs
      b) Barrier reefs
      c) Atolls (Fig. 24)

by the writer to devise a scheme satisfactory for instructional purposes. Over a number of years repeated revisions have been made, but there is undoubtedly further need of change before a thoroughly workable system can be set forth.

DISCUSSION OF THE NEW CLASSIFICATION

Omissions.—Possibly the first reaction of students of shorelines to the accompanying classification will be that important groupings have been omitted. It has already been explained why it was inadvisable to include shorelines of emergence and of submergence in the classification, but the absence of such groups as "plains of marine deposition" or "plains of marine erosion" might be questioned. It is quite conceivable that a relatively straight coast could owe its straightness to recent uplift exposing the even-sloping surface of marine erosion or marine deposition. However, the complicating effects of recent sea-level changes have made it difficult to recognize such surfaces because of their truncation by streams during low sea-level stages. Furthermore, where marine plains of erosion or deposition have been uplifted since the ice melting stopped, they have become quickly converted into "secondary" coasts. This is well illustrated by the irregular wave-eroded pattern of many of the uplifted terraces along the California coast. Also, the offshore bars along the Gulf Coast and the current-straightened east coast of Florida are indicative of the modification of coasts which are at least supposed to have emerged in relatively recent times. Possibly "fold coasts" could have been included, although this type is very difficult to recognize and no examples were known.

Ria coasts.—The first two subdivisions under the drowned river valley classification include coasts where rivers have cut valleys which are now partly flooded by the ocean. The resulting estuaries are with few exceptions shallow and in this way contrast with the estuaries of glaciated coasts. The shape of the estuaries is of course

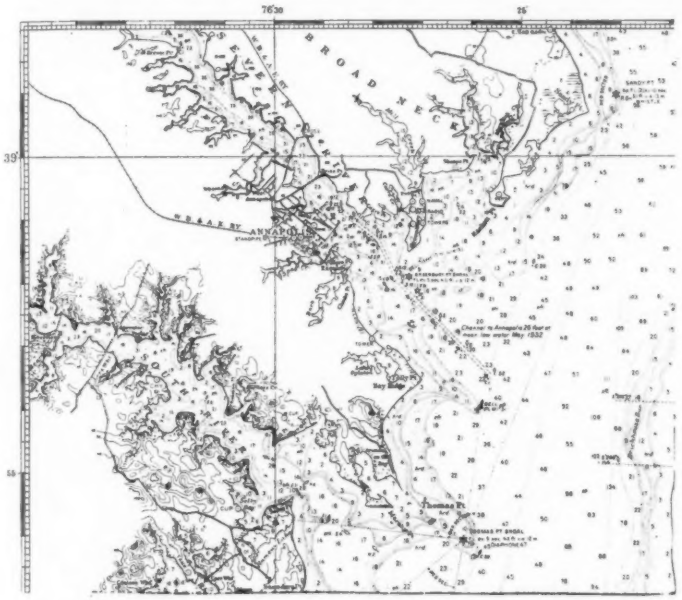

FIG. 3.—A portion of Chesapeake Bay illustrating the dendritic estuary type of coast. (Coast and Geodetic Survey Chart 1225.)

dependent on the structure of the rocks which developed the pattern of the stream courses. Figures 2 and 3 illustrate examples of these rias in horizontal and folded rocks respectively.

Fiord and glacial trough coasts.—Where glaciers advanced into the sea during the glacial period they seem to have almost invariably cut the pre-existing valleys below sea-level. The resulting depressions which have now been flooded by the returning ocean differ from the ria type of estuary in the same ways in which glacial valleys differ from river valleys. Namely, these glacial excavations have

straighter sides, trough shapes with hanging valleys on the side, much greater depths commonly exceeding 100 fathoms, and numerous deep rimmed depressions. There are all gradations between the narrow, steep-walled fiords and the broader, less precipitous walled troughs. The troughs have been described in a previous publication.[7] They are represented by such bays as the Gulf of St. Lawrence

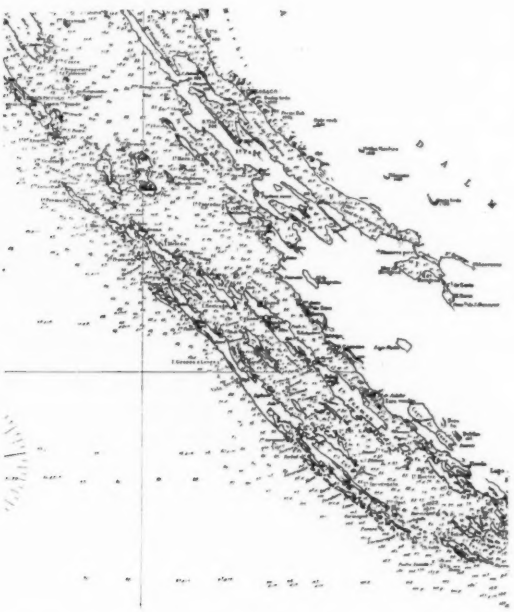

FIG. 4.—Angular or trellis pattern estuaries on the Dalmatian coast. (Hydrographic Office Chart 3957.)

(Fig. 6), the Bay of Fundy, and the White Sea. The fiords (Fig. 5) occur along most glaciated coasts and are well known from their examples in Norway, Scotland, Alaska, and Patagonia. The fiord coasts of Newfoundland and of Maine have not been as commonly recognized as such, partly because the bulk of their relief is submerged.

[7] F. P. Shepard, "Glacial Troughs of the Continental Shelves," *Jour. Geol.*, Vol. XXXIX (1931), pp. 345–60.

River deposition coasts.—The delta type of coast is easily recognized on a chart or map because of its lobate bulge. As explained

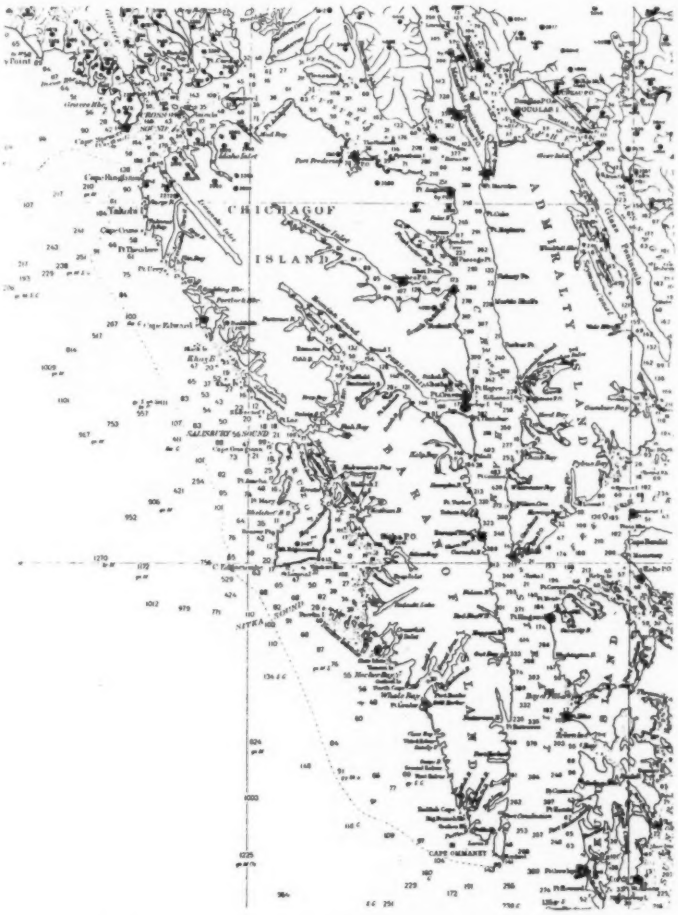

FIG. 5.—A fiord coast, southeast Alaska. (Coast and Geodetic Survey Chart 8002)

by Johnson,[8] there are several types of deltas which include the lobate, the cuspate, and arcuate varieties. These deltas and their

[8] D. W. Johnson, *op. cit.*, pp.187–88.

shapes are of course in part the result of the action of the waves and currents on the deposits of streams entering bodies of standing water. Where alluvial plains are being built out into the ocean (Fig. 7) the waves and currents must also have some straightening effect on the coasts since they would probably be somewhat lobate if the currents were not operating. Thus these river deposition coasts are somewhat intermediate between "primary" and "secondary" coasts, but they fit more with the "primary" group.

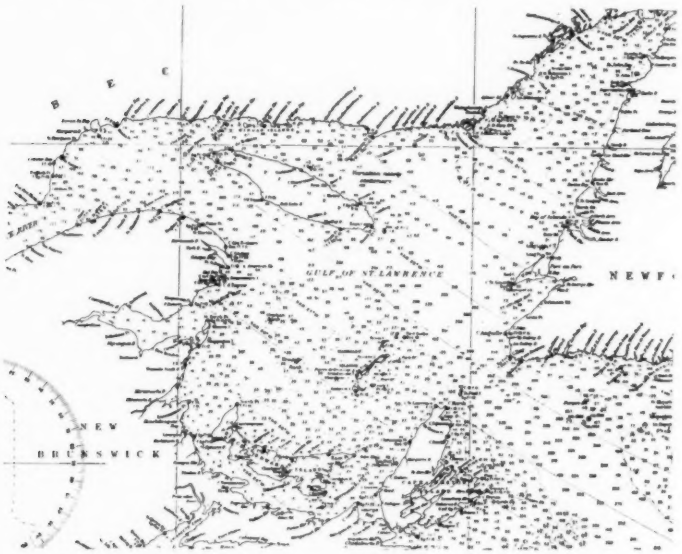

FIG. 6.—The Gulf of St. Lawrence, a glacial-trough coast. (Hydrographic Office Chart 1412.)

Glacial deposition coasts.—There are not very many cases where the coast may be said to be primarily due to glacial deposition. Part of the coast of Long Island represents the front of a moraine so that it might be included in this category (Fig. 8) although the waves and currents have played an important part in its development. Partially drowned drumlins are well illustrated in the case of the islands and of part of the mainland in the vicinity of Boston Harbor (Fig. 9), but these also have been subject to modification by the waves.

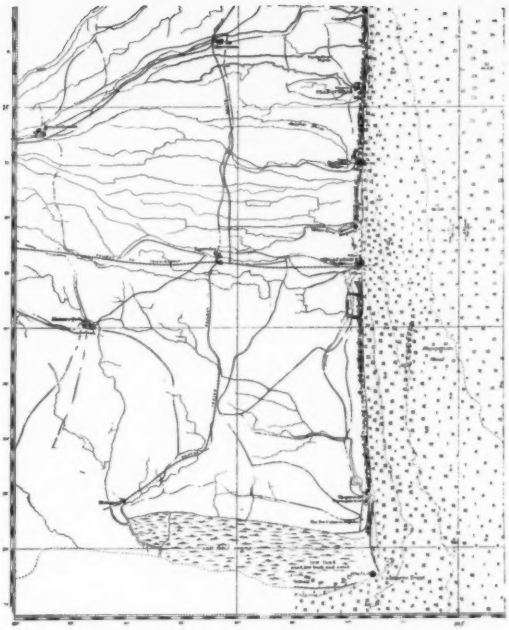

FIG. 7.—An alluvial-plain coast, southeast India. (Hydrographic Office Chart 2434)

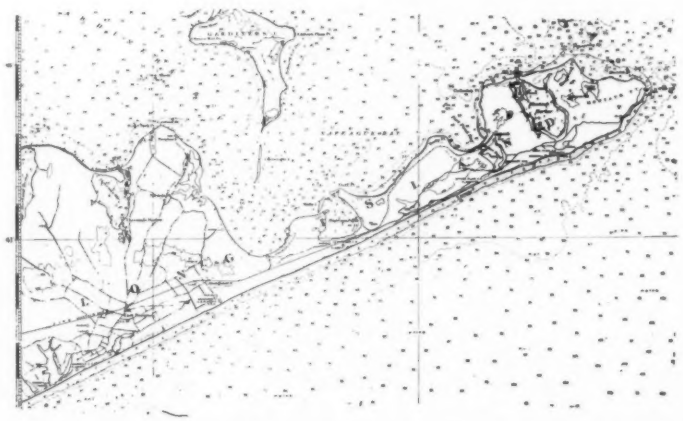

FIG. 8.—A partially submerged moraine, eastern Long Island. (Coast and Geodetic Survey Chart 1211.)

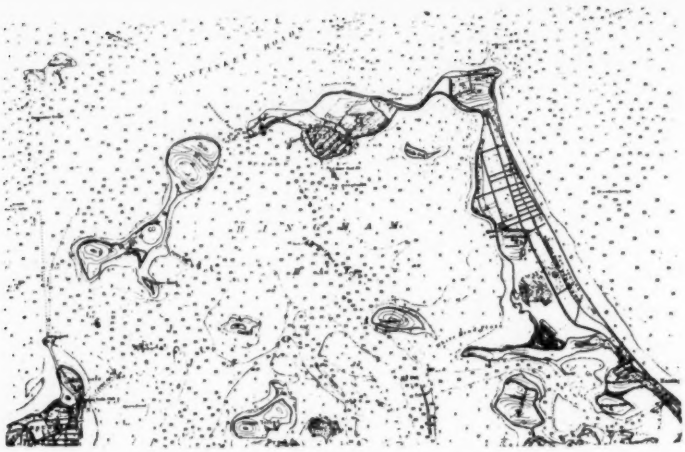

FIG. 9.—Partially submerged drumlins making up Boston Harbor. (Coast and Geodetic Survey Chart 246.)

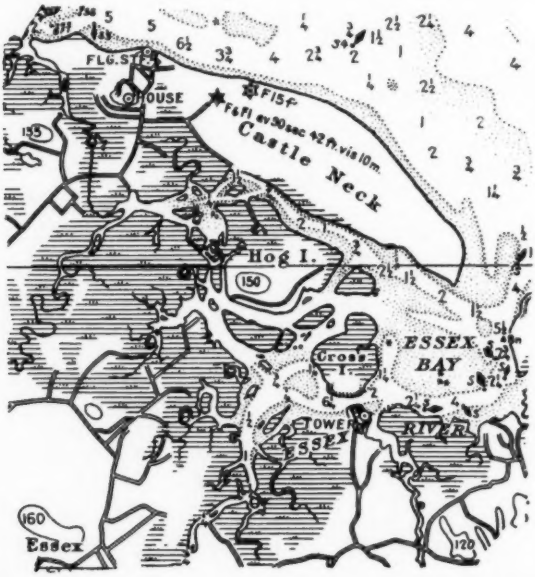

FIG. 10.—A coast built out by wind deposition, northern coast of Massachusetts. Castle Neck has been extended. (Coast and Geodetic Survey Chart 1206.)

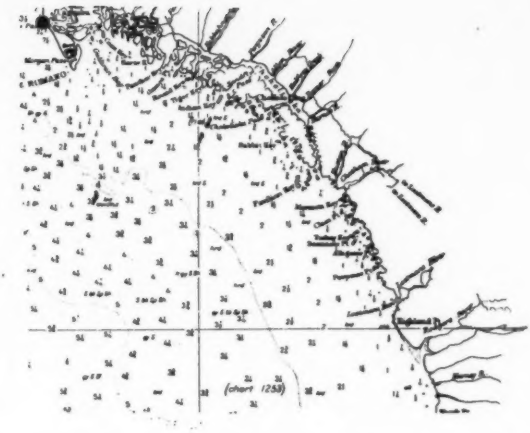

FIG. 11.—A mangrove coast, southwest Florida. (Coast and Geodetic Survey Chart 1113.)

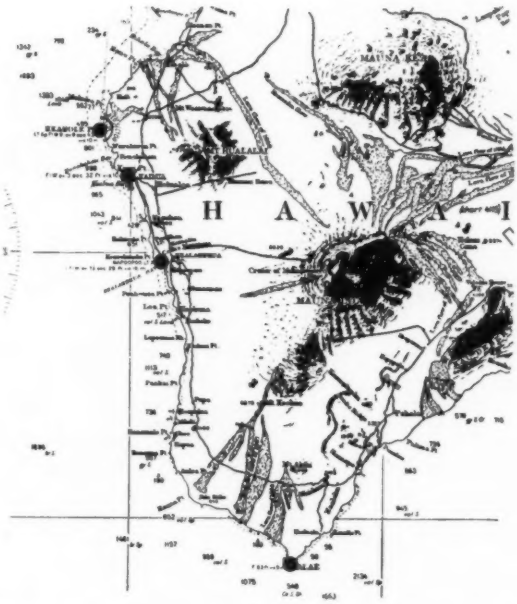

FIG. 12.—Lava-flow coast, Hawaiian Islands. (Coast and Geodetic Survey Chart 4115.)

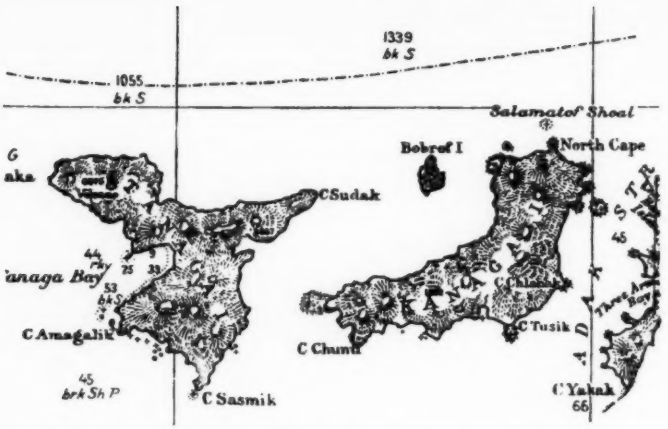

FIG. 13.—Volcanic-explosion coast, Aleutian Islands. (Coast and Geodetic Survey Chart 9102.)

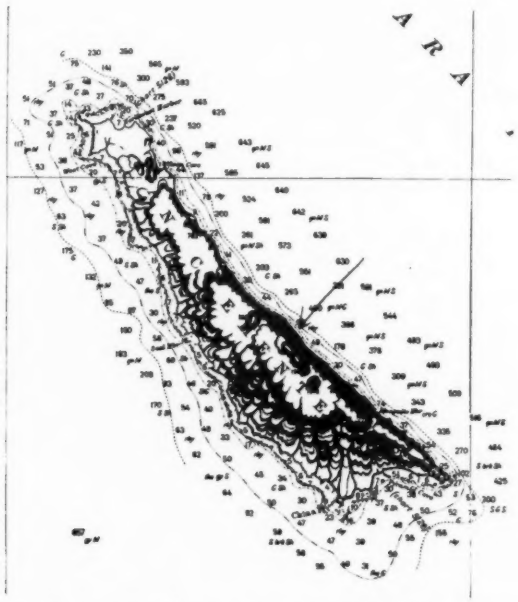

FIG. 14.—Fault-scarp coast, illustrated by the northeast side of San Clemente Island off the California coast. Note absence of shelf. (Coast and Geodetic Survey Chart 5102.)

Wind deposition coasts.—As in the case of "glacial deposition coasts," there are few cases which can be said to owe their form to wind deposition. However, in some places sand dunes are advancing the coast in the direction of the prevailing winds. These coasts are generally of small extent and generally found on the leeward side of a sandy neck of land (Fig. 10).

Volcanic coasts.—Wherever volcanic activity has been comparatively recent in its effects, the coastal forms are dependent on this process. Typical volcanic lava coasts are convex outward as the great lava flows of western Hawaii well illustrate (Fig. 12). On the other hand the concavity of the inside of a partially exploded island is a well-known feature (Fig. 13).

Fault coasts.—In many cases where recent faulting has left fault scarps along the coast the continental shelves have been greatly deepened or eliminated. Therefore a good criterion for the recognition of a "fault scarp coast" would be absence of a shelf (Fig. 14). On the other hand, a "fault line coast" would generally have a well developed shelf which was the result of wave erosion into the soft formations which make up the seaward wall of the fault (Fig. 18). Both coasts would be relative-

FIG. 15.—Fault-trough coast, Gulf of California. (Hydrographic Office Chart 1006.)

ly straight. Other criteria for the recognition of the fault scarps and fault line scarps are the same as those for the lands which have been described by various authors.

There are a number of coasts which on the basis of the chart characteristics seem to fall under the classification of "fault scarp

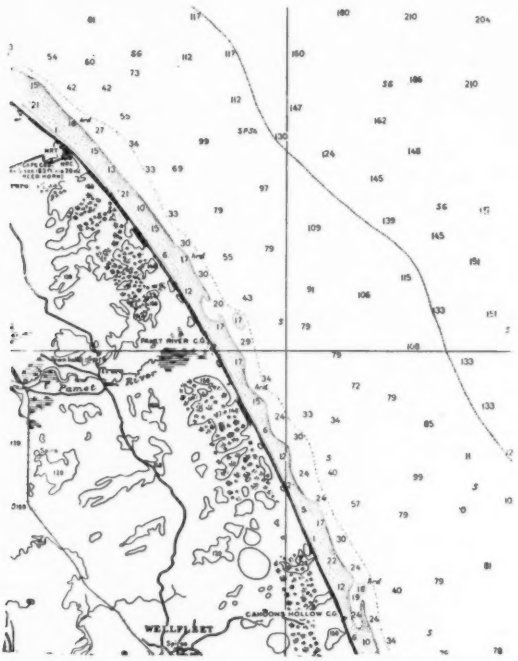

FIG. 16.—Straightened seacoast in homogeneous material, east coast of Cape Cod. (Coast and Geodetic Survey Chart 1208.)

coasts." The most notable examples are along the east coast of Africa, particularly the long straight stretch of Italian Somaliland where there is practically no shelf at all. Also much of the west coast of South America is quite clearly a fault scarp, although in some places narrow shelves have been cut into this escarpment. Also the coast south of Point Sur, California, along the west side of the Santa

Lucia Range has the character of a fault scarp. Other cases are somewhat less certain and need not be mentioned.

The coasts with fault troughs and narrow arms of the sea entering the land are typified by the Gulf of California (Fig. 15) and the

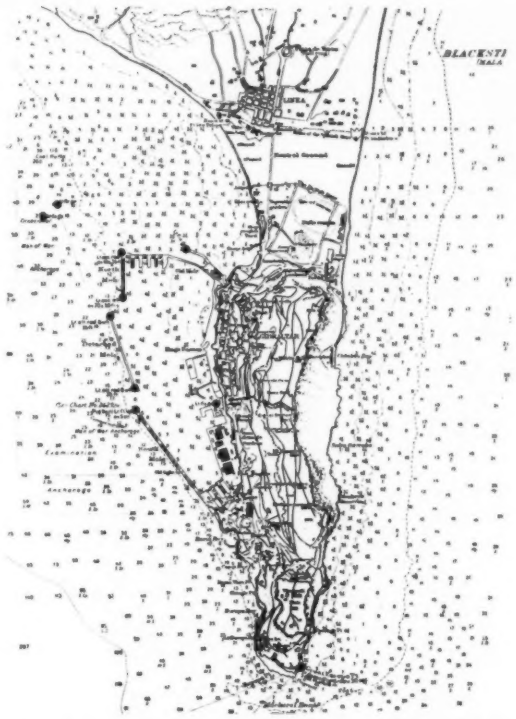

FIG. 17.—A hogback coast, the Rock of Gibraltar. (Hydrographic Office Chart 2430.)

Red Sea. On the other hand, cases like the Juan de Fuca Straits and the Gulf of St. Lawrence, which have sometimes been referred to as fault troughs, appear to be explained better as glacial troughs.[9]

[9] F. P. Shepard, "St. Lawrence (Cabot Strait) Submarine Trough," *Geol. Soc. Amer. Bull. 42* (1931), pp. 853–64.

Wave-straightened coasts.—A point which has been too commonly overlooked by those who wished to classify coasts as emergent on the basis of straightness is that the waves are quite capable of straightening coasts in relatively short periods. The conditions where wave erosion is favorable to straightening are found where

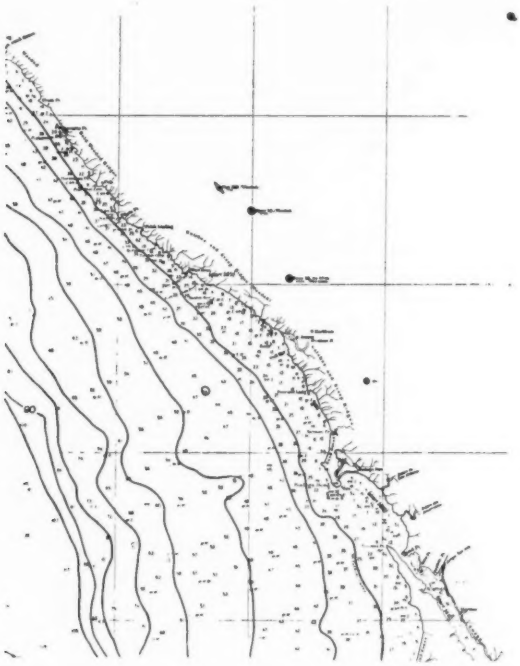

FIG. 18.—A fault-line coast, following the San Andreas fault north of San Francisco. (Coast and Geodetic Survey Chart 5502.)

the coasts consist of uniform materials (Fig. 16) either consolidated or unconsolidated, or where hard layers run parallel to the coasts and are flanked by soft material. The latter condition may develop either in the case of folded rocks where a "hog back coast" will result (Fig. 17) or where there is an old fault with more resistant material on the inside making a "fault line coast" (Fig. 18).

Jagged wave-eroded coasts.—Wave erosion causes irregular coasts where rocks of uneven hardness are attacked. Furthermore, where

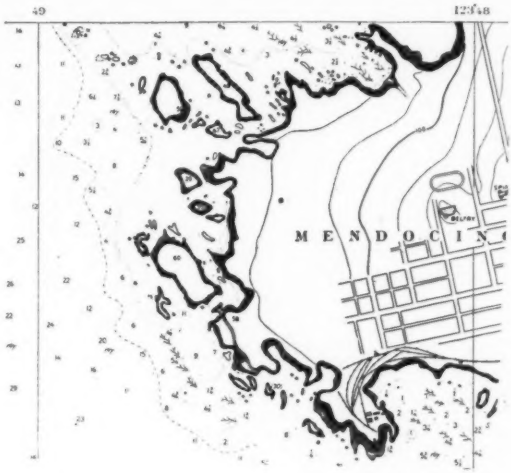

FIG. 19.—Coast made irregular by wave erosion, northern California. (Coast and Geodetic Survey Chart 5710.)

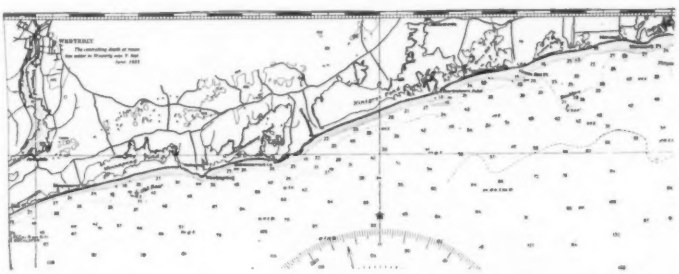

FIG. 20.—Coast straightened by building of bars across estuaries, coast of Connecticut. (Coast and Geodetic Survey Chart 1211.)

low uplifted terraces are being vigorously eroded by the waves, the coasts are commonly jagged regardless of the nature of the rock or materials attacked. It seems possible that the irregularity represents an early stage of development and that it would subsequently

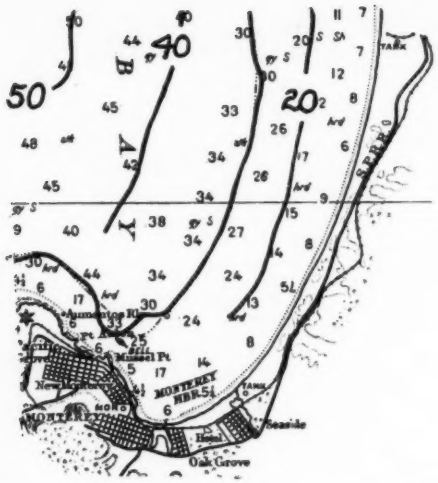

FIG. 21.—Coast prograded by wave deposition, Monterey, California. (Coast and Geodetic Survey Chart 5402.)

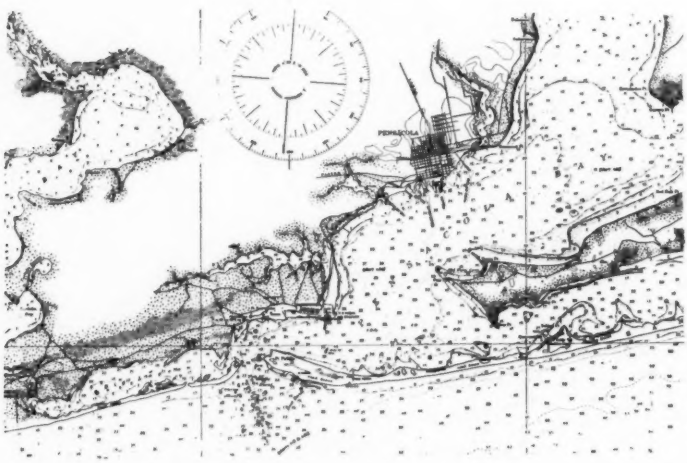

FIG. 22.—Coast with offshore bar and longshore spit, Pensacola, Florida. (Coast and Geodetic Survey Chart 1265.)

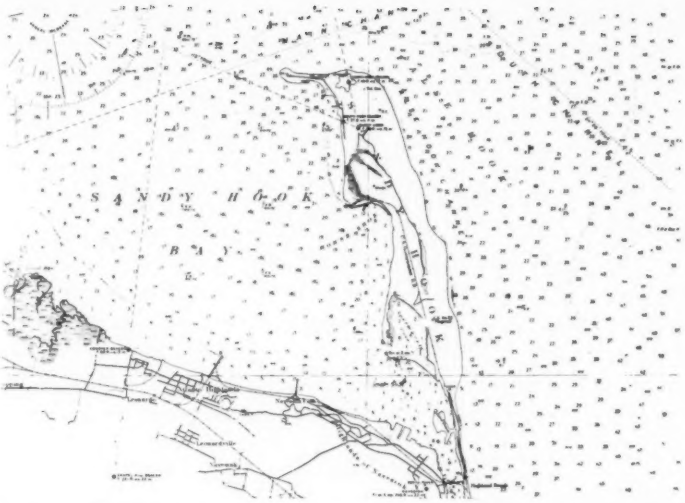

FIG. 23.—Sandy Hook, a longshore spit. (Coast and Geodetic Survey Chart 369.)

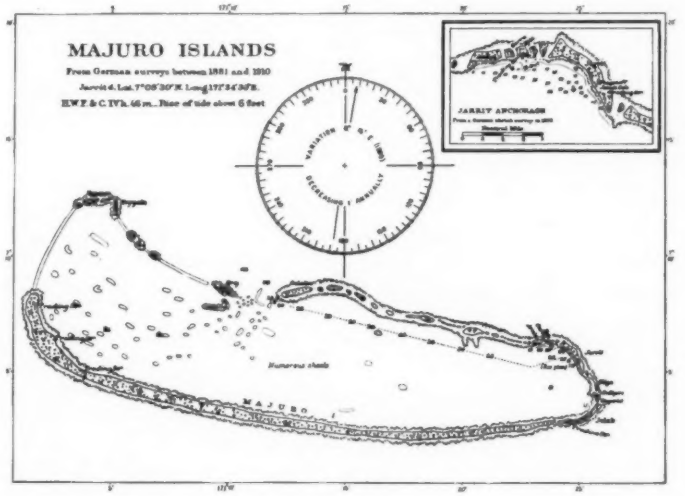
MAJURO ISLANDS
From German surveys between 1881 and 1910
Jarvis d. Lat. 7°08'30"N. Long 171°24'30"E.
H.W.F. & C. IV h. 46 m. Rise of tide about 6 feet
JABBIT ARCHIPELAGO
Numerous shoals
MAJURO

FIG. 24.—An atoll, Marshall Island, North Pacific Ocean. (Hydrographic Office Chart 2031.)

be superseded by a straightening, particularly where the material attacked is comparatively uniform. The jagged coast is commonly bordered by numerous stacks (Fig. 19).

Coasts with bars and spits.—Very early in the evolution of an indented coast, as Johnson has pointed out, spits start to build across the entrance of the estuaries. Unless tidal currents or violent wave action interferes, these spits presently develop into bars. In this way the coast may become straightened by deposition (Fig. 20).

Another type of deposition occurs where the coast extends with a very gentle slope out under the water. Here offshore bars are built parallel to the shore but at distances ranging up to 10 miles or more from the mainland. These bars extend for many miles along low coasts. Most of them are connected at some point to the mainland (Fig. 22). Coasts of this type are particularly common in southeastern and southern United States. As explained previously, they are not necessarily the result of emergence of the coast in recent times.

In some places at the end of these offshore bars there are great spits with curving ends developing hook shapes which make important protection for harbors. Sandy Hook is a particularly good example (Fig. 23).

Other types.—In the preceding discussion some of the classes of coasts have not been mentioned. These omissions were made because the types are so well known that their discussion seemed superfluous.

HYPOGENE EXFOLIATION IN ROCK MASSES

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ABSTRACT

Exfoliated rock masses have been observed in mine workings below the zone in which atmospheric agencies and insolation operated. Exfoliation is found in rock masses, of radii ranging from one-quarter inch to thousands of feet, under diverse geologic environments. Only one agency appears to have been operative in all cases of exfoliation—dilation following the unloading of earth pressure. In special cases of rock exfoliation surficial agencies may have been auxiliary causes, enhancing the dilation.

INTRODUCTION

A discussion of exfoliation in rock masses has been included in nearly all texts of physical and general geology for the last half-century. Exfoliation had not been given much individual treatment until Eliot Blackwelder[1] rejected insolation as an important agent in the development of the structure. He later summarized the ideas involved[2] and ably surveyed the literature pertaining to exfoliation; the reader is referred to this work for general treatment and for bibliography. Laboratory experiments[3] have recently supported Blackwelder's conclusion that insolation is not important in the development of exfoliation.

The currently favored interpretation is that exfoliation in rocks is caused chiefly by volume changes resulting from chemical attack on the rock by moisture during weathering. A distinction has been drawn[4] between exfoliation, considered a mechanical process, and spheroidal weathering, considered a chemical process. This generic classification is desirable but cannot be put in use while doubt lingers

[1] "Exfoliation as a Phase of Rock Weathering," *Jour. Geol.*, Vol. XXXIII (1925), pp. 793–806.

[2] "The Insolation Hypothesis of Rock Weathering," *Amer. Jour. Sci.*, Vol. XXVI (1933), pp. 97–113.

[3] David T. Griggs, "The Factor of Fatigue in Rock Exfoliation," *Jour. Geol.*, Vol. XLIV (1936), pp. 783–96.

[4] W. J. Miller, *An Introduction to Physical Geology* (New York: D. Van Nostrand Co., 1925), p. 68.

about the origin of the structures. Exfoliation is a structural term used here to designate all masses made up of concentric shells of rock separated by definite, smoothly curved partings.

HYPOGENE EXFOLIATION IN GRANODIORITE, AMADOR COUNTY, CALIFORNIA

Geologic setting.—The Sierra Nevada batholith, of Jurassic age,[5] is cut along its western rim by pre-Tertiary gold-quartz veins, called the East Belt deposits. They are located 15 miles east of the Mother Lode district at the latitude of Jackson, California. Portions of the vein-bearing granodiorite are capped by andesitic lava of probable Tertiary age. In the Defender mine andesite-porphyry dikes traverse the granodiorite and cut cleanly through the quartz vein. The dikes probably were the vents from which some of the overlying lava was extruded.

The granodiorite country rock near the vein has been definitely altered by solutions accompanying the vein matter. The most characteristic of the alteration products is sericite (possibly pyrophyllite) which has been developed in small masses and fringes along the vein walls and in the near-by fractured granodiorite. The dikes are not accompanied by visible sericite, nor is there any other likely source for it than the mesothermal gold-quartz vein.

Exfoliated boulders.—On the portion of the Defender area which is not capped by lava many rounded boulders of granodiorite have resisted weathering and stand up boldly above the weathered surface. Some of the boulders are typically exfoliated. They would be commonplace if it were not that in a near-by mine working, about 50 feet below the lava capping, the slightly weathered granodiorite contains several masses made up of concentric rock shells, of one- to four-foot radii, separated by smoothly curved partings. Exfoliation is developed more perfectly in these subsurface masses than in the boulders that outcrop in the area. The critical feature of the subsurface exfoliation is a thin layer of sericite which is plastered to the granodiorite along some of the curved partings. The sericite appears similar to that found elsewhere near the quartz veins and, as no other

[5] George W. Stose, *Geologic Map of the United States* (Washington: U.S. Geol. Survey, 1932).

source of the mineral could be found, it must be concluded that the sericite was introduced into the partings at the time the veins were emplaced.

HYPOGENE EXFOLIATION IN "PEBBLE DIKES"

The "pebble dikes."—In the Tintic mining district, about 60 miles south of Salt Lake City, Utah, are found exposures of the interesting deposits known as "pebble dikes" and the larger breccia pipes of similar character. They are lenses or chimneys of mud matrix surrounding a mass of rounded to subangular fragments of country rock which were torn from the sides of the conduit and were carried thousands of feet upward to the loci of deposition by solutions of magmatic origin under strong intrusive pressure.

The "dikes" have been described in detail[6] in reference to their association with ore deposition. They are of importance here because many of the constituent rounded "pebbles" exhibit well-developed exfoliated structure (onion-structure) which is lacking in all ordinary boulders derived from the same rock formations, either where found in weathered outcrops or in the long-weathered Tertiary river gravels (Fig. 1).

Hypogene exfoliation in "pebbles."—The exfoliated "pebbles" are chiefly of the Tintic formation, a pure, close-grained Cambrian quartzite which is the basal member of the sedimentary rocks exposed in the district. The quartzite is homogeneous except for the few bandings parallel to bedding which are without influence on the curved exfoliation partings. The limestone "pebbles" are rarely exfoliated at Tintic but are abundantly exfoliated in the near-by Park City district.

Mine openings at Tintic have exposed "dikes" bearing the exfoliated "pebbles" through a vertical range of 4,000 feet, in many places far below the Tertiary and present zones of weathering. The abundant surface and underground exposures of the "dikes," coupled with the highly characteristic stratigraphy of the district, give an unusual opportunity to study the transport of the "pebbles" from their parent formations to the points of deposition. Many of

[6] Rollin Farmin, "Pebble Dikes and Associated Mineralization at Tintic, Utah," *Econ. Geol.*, Vol. XXIX (1934), pp. 356–69.

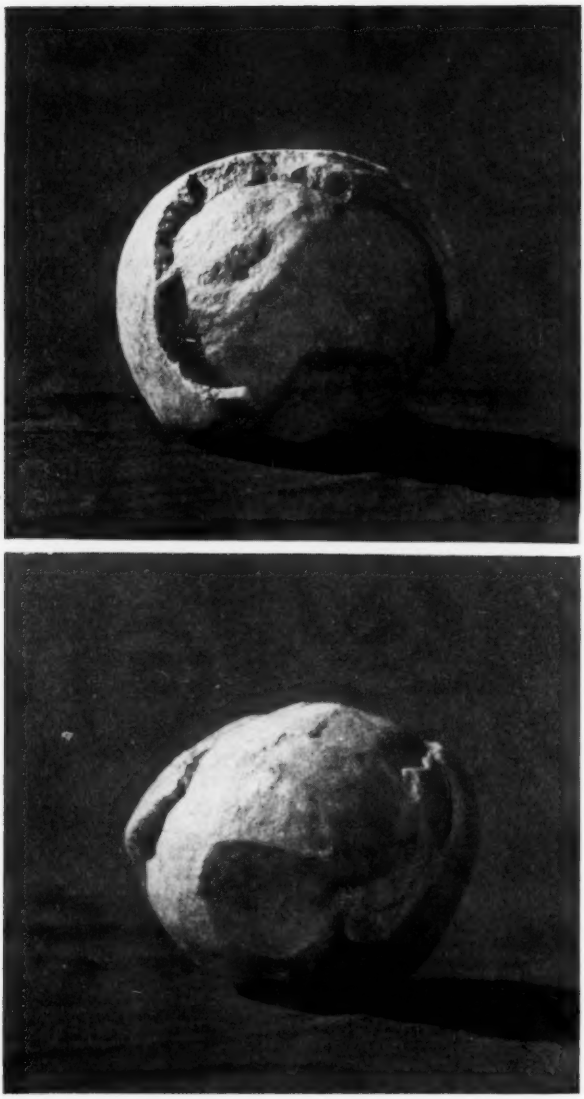

FIG. 1.—Above: two-inch exfoliated "pebble" of quartzite from a "pebble dike," Tintic district, Utah. Below: the opposite side of the "pebble," showing well developed onion-skin structure.

the exfoliated "pebbles" have had segments of their outer shells peeled away during transport because the missing segments cannot be found in the matrix immediately surrounding them, although random segments of the shells are sparingly found throughout the matrix (Fig. 2).

FIG. 2.—Segment of quartzite shells found in a "pebble dike" in Sonoma Range, Nevada. The segment has been peeled from a large "pebble." Loaned by Dr. Siemon W. Muller.

Further evidence which helps to place the time of development of the exfoliation is: "Pebbles" found between walls of their parent formation—and which therefore have only been transported a short distance—are commonly several inches in diameter, and the discernible exfoliation is limited to the outer part of the spheres. In contrast, quartzite "pebbles," carried upward into a formation lying 6,000 feet stratigraphically higher than the Tintic, have been worn to one-tenth the diameter of those found closer to their source, but they are nonetheless exfoliated. The exfoliation must

have been developing progressively while the "pebbles" were carried upward toward the surface.

THEORIES OF ORIGIN OF EXFOLIATION

Insolation theory.—As has been noted above, insolation has been rejected as a significant agency in rock exfoliation by recent workers. The field evidence of exfoliation in hypogene environment completely rules out insolation as a factor in the cases at hand. The theory should be discarded until an authentic case of exfoliation can be attributed to it.

Theory of chemical attack.—In considering the origin of rock exfoliation we may look about for similar structures developed in other materials. At Grass Valley, California, in a reopened mine adit, long sealed by a cave-in at its portal, I recently found a broken drill steel from which shells of iron oxide one-eighth of an inch thick were exfoliated. The steel had been in moist air, at constant temperature 90 feet below the surface, for forty-five years. Its exfoliated structure was developed by chemical attack alone; the partings separate iron oxide from steel and from other layers of oxide but do not penetrate into the steel core.

Similar exfoliation may be expected to develop in rock masses as a result of chemical attack during weathering. Rock masses on the surface are commonly only slightly weathered, and the difference in degree of weathering between shells of exfoliation may be imperceptible. The more completely weathered granite surrounding boulders of exfoliated rock does not contain concentric fractures; consequently there is little parallel between the proved case of chemical exfoliation in steel and exfoliation in weathered rocks.

The theory of exfoliation by chemical attack should be competent to explain both surface and hypogene exfoliation if it is to be considered the sole cause of the structure. Weathering is not a factor in the exfoliation found in "pebble dikes," and any hypogene chemical attack upon the "pebbles" left no trace which could be seen with the petrographic microscope. "Pebbles" of sandy limestone should show differential leaching but none can be found. The assorted "pebbles" of pure quartzite, shaly quartzite, shale, sandy and shaly limestone, and pure limestone present an ideal background against which to

view the attack of solutions, either supergene or hypogene. Furthermore, it would be a remarkable coincidence if the chemical action of both types of solutions were to generate concentric stresses in rocks of diverse chemical compositions. The theory of chemical attack is inadequate to explain all types of exfoliation.

Theory of exfoliation concentric around nuclei of solidification.—A very perfect exfoliate structure is developed by freezing a hard-boiled egg. Thin septa of ice separate the white of the egg into a dozen symmetrical shells which, after the ice has melted away, markedly resemble the concentric shells of an exfoliated boulder. The structure is not entirely developed by the ice wedges because one or more partings of the same type can be found in the cooked egg before it is frozen; moreover, no such structure is found in an uncooked egg that has been frozen. This series of homely experiments leads me to conclude that the structure is developed during the solidification of the mass and is only magnified by the freezing.

By transferring the foregoing theory to exfoliation in rock masses, it would seem possible that the structure might center about nuclei of solidification of the granite from the magma. Invisible partings or strains in the rock developed during consolidation might be greatly magnified by the action of weathering agencies or by the action of hypogene solutions. In conflict with this theory is an exfoliated mass of granodiorite at Nevada City, California, where an autolith several inches in diameter is cut cleanly through by several of the outer concentric partings, with slight fault offsets of the "contact." The autolith presumably represents a point of early (or late) solidification from the magma, but, because it is not centrally located in the exfoliated mass, the theory is weakened.

Another conflict with the solidification theory is the exfoliation developed in limestone and quartzite "pebbles"—rocks of sedimentary origin which have not consolidated in concentric form.

Theory of dilation after unloading of earth pressure.—Observers of exfoliation have agreed that increase in volume is the factor in the process which results in the concentric fractures in a rock mass. When the volume increase has developed stresses that exceed the elastic strength of the rock concentric fractures result. In consideration of the field evidence the theories of insolation and chemical

attack have been rejected because neither will explain the diverse cases of exfoliation. A force remains which could have been operative in all cases of rock exfoliation and which has demonstrated its ability to rupture rocks—dilation after unloading of earth pressure. This agency was suggested by G. K. Gilbert[7] in explanation of dome structure.

Rock bursts are a familiar hazard in deep mines where man has removed support from one side of a rock mass, and they demonstrate with fatal violence the existence of earth pressure which exceeds the elastic strength of unloaded rock. The pressure may involve factors other than hydrostatic and lithostatic load—for example, the carbon-dioxide eruptions in coal mines.[8]

Rock bursts in mines are of two general types: The first is the collapse, owing to superincumbent load, of the hanging wall at wide spans in stopes, accompanied by the crushing of rock pillars. The second is the explosive ejection of small rock fragments from rock surfaces in narrow, newly opened mine headings where there is no general failure of the ground. In the second type of burst the internal expansive stress overcomes the tensile strength of the unloaded rock.

The unloading of rock masses is continually in progress as erosion brings nearer and nearer to the surface rocks adjusted to the heavy pressure which must prevail in the deeper portions of the earth's crust. The unloaded rock will tend to expand and will do so by fracture wherever internal stress exceeds its elastic strength. Not all of the fractures will be concentric, but a concentric exfoliation is the ideal form of relief of the stress in a homogeneous rock. In plutonic igneous rocks, at the time of crystallization, the mineral grains were bonded together under heavy pressure. Similarly the bond between mineral grains of deeply buried or folded sedimentary and volcanic rocks will tend to become adjusted to heavy load. When erosion brings a rock of this type to the surface a dilation of the rock will tend to ensue, proportionate to the unloading. The dilation may

[7] "Domes and Dome Structure of the High Sierra," *Geol. Soc. Amer. Bull. 15* (1904), pp. 29–36.

[8] Otto Stutzer, "Carbon Dioxide Eruptions from Coal Seams in Lower Silesia," *Econ. Geol.*, Vol. XXXI (1936), pp. 441–52.

cause doming and exfoliation, independent of weathering. As the rock is weakened by the subsequent attack of weathering, a further exfoliation may develop at points where the rock was strong enough to withstand its internal stress before attack.

Rocks bursts are commonly experienced in mines at the depth of 5,000 feet and in some mines at half that depth. Fully 10,000 feet of

FIG. 3.—Exfoliated boulders and domes of granodiorite of assorted sizes, Nevada City, California. Left: four-foot exfoliated boulder very similar in form to the "pebble" in the upper illustration of Figure 1. Right: whale-backed exfoliated mass, 20 feet long.

rock are believed to have been eroded from the Sierra Nevada during Tertiary time; hence the relief of load from rocks at the present surface has been of a magnitude which could produce violent rock fracture. F. D. Adams[9] produced a pseudo-exfoliated structure in spheres of marble compressed in a viscous medium under 100 tons pressure. His experiment dealt with compression, however, rather than with the dilation after compression.

Whale-backed and spheroidal rock masses exhibiting slight ex-

[9] "An Experimental Investigation into the Action of Differential Pressure on Certain Minerals and Rocks," *Jour. Geol.*, Vol. XVIII (1910), Pl. Vc, p. 517.

foliation, of sizes intermediate between large domes and the smaller exfoliated boulders, are common in the Sierra Nevada and elsewhere (Figs. 3 and 4). They complete a series of similar structures ranging in size from "pebbles" of one-quarter-inch diameter to the mighty

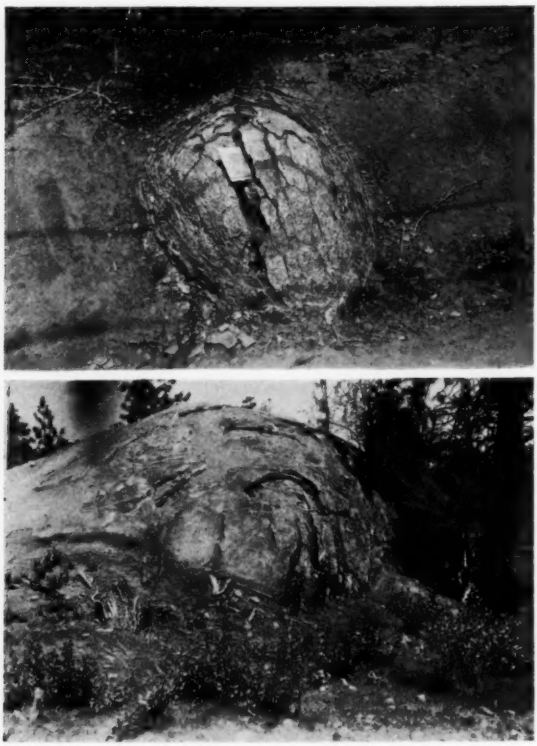

FIG. 4.—Exfoliated boulders and domes of granodiorite of assorted sizes, Nevada City, California. Above: exfoliated boulder of eight-foot diameter. Below: a small dome, markedly exfoliated.

domes of the Yosemite. All are marked by curved partings of unusual symmetry, developed more or less concentrically. The onion-skin "pebbles" at Tintic were carried upward thousands of feet toward the surface; during the journey earth pressure upon them was

continually reduced and they progressively developed exfoliated structure. The larger forms of exfoliated rock masses are likewise best explained by the theory of dilation after unloading of earth pressure, which is the only agency which appears to have been operative in all cases of exfoliation.[10]

SUMMARY AND CONCLUSIONS

Hypogene exfoliation in rock masses has been observed in mine workings in Amador County, California, which exhibit features which cannot be explained by the theories of insolation or of chemical attack during weathering. Exfoliation in the "pebble dikes" at Tintic, Utah, was also developed without the aid of surficial agencies. Among the several theories of exfoliation considered, only that of the dilation of the rock masses after unloading of heavy earth pressure seems adequate to explain the diverse examples of the structure, which range from the onion-skin "pebbles" of one-quarter-inch diameter to the domes of the Sierra Nevada. Many surficial processes may be co-operative with unloading in the development of rock exfoliation, but they are probably of secondary importance.

[10] As this paper goes to press I learn that at the recent thirty-sixth meeting of the Geological Society of America François E. Matthes read "Exfoliation of Massive Granite in the Sierra Nevada of California," in which he reports exfoliated structure in U-shaped glacial valleys, where it is developed on huge scale parallel to the valley floors and walls. He interprets the structure as caused by dilation of the granite following unloading brought about by the rapid glacial erosion.

EVIDENCE OF MULTIPLE GLACIATION IN THE NORTHERN PART OF YELLOWSTONE NATIONAL PARK

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ABSTRACT

Multiple glaciation is evident in the north-central part of Yellowstone National Park; probably the youngest Pleistocene is Wisconsin in age. The valley formerly occupied by Lava Creek is pre-Wisconsin in age, and it has undergone little modification since that time. The valley contains old, deeply weathered granitic glacial boulders overlain by patches of fresh drift. During early Pleistocene time, Lava Creek occupied the valley between the Gardiner River and Blacktail Deer Creek and drained northward into the Yellowstone River. It has abandoned the pre-Wisconsin channel and drains into the Gardiner River at the west end of the old valley. This change was brought about by an ice lobe from the Lamar River region, probably in late Pleistocene time.

INTRODUCTION

The glacial history of Yellowstone National Park has received the attention of geologists for the past sixty-five years. For the most part, however, the work has been of a general nature in connection with the interpretation of the various features present. Detailed glacial studies have been almost entirely lacking and multiple glaciation, though generally assumed, had not been demonstrated.

Two field seasons in the Park area, together with a study of the literature pertaining to its glacial history, convinced the writer of the desirability of a detailed regional study within the confines of the Park. As a "key" area he chose the north-central portion of the Park (Fig. 1), which lies between parallels 44°55″ and 54°02″ N. and meridians 110°30″ and 110°50″ W.

The writer is grateful to Dean George F. Kay of the State University of Iowa who has given freely of his time and counsel in the preparation of this manuscript and the work on which it is based. During the summer of 1936 Dean Kay went over the area with the writer and offered many valuable suggestions. To the National Park Service the writer is deeply indebted for the use of the Park Service facilities. Dr. C. Max Bauer, chief naturalist, read the entire manuscript and gave untiringly of his counsel and the facilities of his

office. W. E. Kearns and the late George Crowe, assistant naturalists, extended many courtesies in the field and laboratory; also Dr. W. C. Alden read the manuscript and made numerous constructive criticisms. For all this assistance the writer acknowledges his indebtedness.

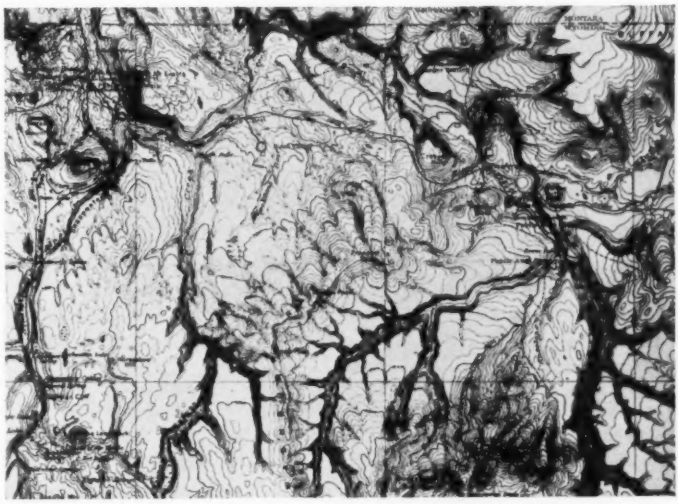

FIG. 1.—Map of the north-central portion of Yellowstone Park. Contour interval, 100 feet. Scale, 1/250,000.

OLD LAVA CREEK VALLEY

In 1878 Holmes[1] visited the region described in this paper as a member of the Hayden Survey. The report which he wrote at that time refers to the Upper Gardiner River, that stream which rises in the region of Little Quadrant Mountain and flows southeast and then northeast between Bunsen Peak and Sheepeater Cliffs (Fig. 1), as the Middle Fork of the Gardiner River. Today this stream is known as Gardiner River. The stream recognized today as Lava Creek, Holmes called the East Fork of the Gardiner River, and the

[1] W. H. Holmes, "Report on the Geology of the Yellowstone National Park," *U.S. Geol. and Geog. Survey of the Territories, 12th Ann. Rept. for 1878*, Part II (1883), pp. 1–57.

east-west valley between the junction of Lava Creek with Gardiner River and Blacktail Deer Creek, the valley of the East Fork. This valley is referred to by the writer as Old Lava Creek Valley. The stream known today as Glen Creek, which flows through Golden Gate, Holmes called the West Fork.

Old Lava Creek Valley is a broad, level, U-shaped valley extending in an east-west direction (Fig. 2). It is about 5 miles long and the floor of the valley is $\frac{1}{4}$–$\frac{1}{2}$ mile wide. Lava Creek enters the valley from the south and turns sharply west, flowing about one mile over the level, basalt-covered, rapidly narrowing valley floor to Undine Falls. It plunges over the brink and drops 60 feet, then cascades 700 feet more to join Gardiner River $1\frac{1}{2}$ miles below the Falls at the foot of the south slope of Mount Everts. Of this valley Holmes wrote:

> There has doubtless been a time when they [the three forks of the Gardiner River] formed a junction before descending from the plateau, and it is not improbable that one fork at least—the eastern—reached the Yellowstone without forming a junction with the other branches by passing down to the east of Mount Everts.[2]

Where Lava Creek enters Old Lava Creek Valley there is a distinct col. The creek turns sharply west on the west side of this col. Glacial drift is relatively thick at the col but thins rapidly eastward. Half a mile east of where the Mammoth Hot Springs–Tower Falls road crosses Lava Creek the valley floor is only thinly mantled with drift. What appears to be an old stream channel carved in the basalt of the valley floor is plainly visible. Continuing eastward down slope from the col, the valley widens rapidly and the drift thickens perceptibly until it becomes exceedingly thick in the eastern end at Blacktail Deer Creek as shown at the extreme right of Figure 3. From a point $2\frac{1}{2}$ miles east in the valley to its debouchure at Blacktail Deer Creek, it exhibits a typical knob and kettle topography. Some of these depressions contain ponds which drain eastward (Fig. 3). Blacktail Deer Creek intersects this valley at right angles. It has cut a deep channel through the drift and drops rapidly to the Yellowstone River 1,200 feet below in a distance of 3 miles (Fig. 3*b*).

[2] *Ibid.*, p. 28.

Photo by Dr. C. Max Bauer, N.P.S.

FIG. 2.—Panorama looking north across old Lava Creek Valley. *a*, former Lava Creek channel; *b*, rhyolite cap on Mount Everts; *c*, recessional moraines.

In passing through the valley one is impressed by the absence of glacial drift on the central part of the valley floor midway in the valley. It is swept relatively clean. Only scattered granite and basalt boulders 2–4 feet in diameter are present. The boulders increase in number toward the valley walls. Some exhibit weathering to a depth of $\frac{1}{2}$ inch.

The sides of the valley are completely covered with a veneer of drift varying in thickness. The highway from Mammoth Hot Springs to Tower Falls on the south side of the valley is constructed

Photo by Dr. C. Max Bauer, N.P.S.

FIG. 3.—Panorama of Blacktail Deer Creek Valley, southeast to Washburn Range. *a*, Old Lava Creek Valley; *b*, Blacktail Deer Creek; *c*, Upper Blacktail Deer Creek; *d*, Washburn Range.

on this material in places nearly 80 feet above the valley floor. The highway cuts through a 15-foot knob of this drift on the east side of the col midway between Lava Creek and Blacktail Deer Creek. The exposure extends for 30 feet where it emerges from the cut only to be exposed again as several successive elevations are penetrated eastward along the road.

The deposit is compact. It is composed of fresh granite and other igneous boulders up to 8 or 10 inches in diameter, many of them striated, in a mass of finer gravel containing considerable clay. No limestone pebbles were found. The boulders are oriented with their long axes parallel with the valley; there is a slight suggestion of sorting and two of the knobs exhibit a gentle easterly dip. The larger boulders appear to be confined to a band about 2 feet wide midway between the top and bottom of the exposure. On the opposite side

of the road nearer to the valley floor where the drift thins, this fresh drift is found to mantle the old, deeply weathered granites. At several places along the highway this condition is found to exist. It apparently indicates a young drift mantling boulders of a previous glacial stage.

The upper foot of the deposit is calcareous. It appears that the calcareous material has been derived secondarily as there is no limestone in the deposit and examination of the drift at a depth of 3 feet and again at 10 feet did not reveal any calcareous material.

Pieces of petrified wood were found on the valley floor and in the young drift on the valley sides. The fragments increase in abundance eastward in the valley.

Where Lava Creek enters the valley and veers west there is a channel leading eastward for 200 yards before its identity is lost (Fig. 2*a*). Today this channel is marked by a swampy strip extending southwest from the highway to form a Y with the present stream bed. A few small flakes of obsidian were found on the valley floor opposite the mouth of this old channel but none was observed in the eastern part of the valley. This fact is quite significant in that Obsidian Cliff, to the southwest at least 10 miles airline, is the only known source of this particular type of obsidian and this valley floor is one of two places east of Lava Creek where fragments of this material were found.

One mile west of Undine Falls a highway cut reveals an exposure of drift 22 feet thick and 100 feet long. The drift is well sorted and there are few if any boulders over 5 inches in diameter. An examination of the material, both in the field and in the laboratory, failed to reveal any pebbles, cobbles, or boulders of sedimentary origin. The nature of the drift (granites, breccia, and particles of petrified wood) indicates it was derived from the east and northeast of the valley.

Less than $\frac{1}{2}$ mile west of this deposit huge blocks of travertine from Terrace Mountain are strung across the valley in a northeast-southwest direction. This narrow strip between the drift deposit just described and the travertine blocks clearly marks the maximum extent of the ice from the east and from the west in Old Lava Creek Valley.

Lava Creek enters the old valley a few feet west of the col and crosses the roadway at the site of a public campground. Opposite the campground on the south side of the road is an exposure of drift 12 feet high and 100 feet long. The deposit is composed of coarse sand and boulders some of which are 3 feet in diameter. The drift is not sorted, but it rests upon basalt that discloses some polish by running water. The basalt is the southeastern continuation of the flow that forms the brink of Undine Falls.

The western end of the old valley, where Lava Creek enters, is in striking contrast to the eastern end. The valley bottom east of the col is broad and level with little drift; the portion west of the col is narrow, with slopes heavily drift-covered, and precipitous to the Gardiner Valley below.

The north side of the valley, embracing the south flank of Mount Everts, has the same gentle slope as the south side of the valley as far west as Undine Falls. From here west to its confluence with the Gardiner Valley the slope is 35°–40°. The crest of Mount Everts, 800 feet above the valley floor, is capped with a rhyolite flow (Fig. 2*b*). The edge of the flow is a 100-foot cliff which overhangs slightly, and it is underlain by volcanic tuff which has been firmly cenented.

At one place a mile west of Undine Falls the rhyolite has apparently been undermined by water for 200 yards and it has slumped. The base of the flow has sagged approximately 15 feet to form a V-shaped trough dipping toward the center at about 10° (Fig. 4). The tuff exposed at other places beneath the rhyolite is gray in color; here it has the pinkish shade of the overlying rhyolite due to the quantities of that material mixed in it. Only the texture and the erosional difference reveal the contact between the two. Holmes observed the volcanic tuff and wrote:

> The most peculiar and interesting circumstance connected with this flow is the existence of a bed of loosely compacted material that lies between the sandstone floor and the sheet of rhyolite. This bed varies in thickness from 5 to 10 feet. The lower part is deposited upon the somewhat uneven surface of the sandstones, and is composed of a white powder that resembles slaked lime in appearance, and it is probably a volcanic tufa.[3]

[3] *Ibid.*, p. 12.

He further states:

> Between the gray laminated layers and the base of the rhyolite is a zone of peculiar rock that seems to show an imperceptible transition from the one to the other. It is quite impossible to tell just where the rock ceases to be an aggregation of grains and becomes a true crystalline lava.

Holmes, however, did not comment upon the slump in the flow at the south end of the mountain.

FIG. 4.—Slump in rhyolite flow, Mount Everts. *a*, break in flow; *b*, drainage terrace; *c*, Lava Creek.

At the closure of the V the rhyolite is fractured and crushed (Fig. 4*a*). It appears that the underlying tuff has been re-worked by water, which has caused angular fragments of the rhyolite to become incorporated in the basal mass. The tuff has been squeezed by the sag until it pinches out rapidly toward the center. Since cementation the lower part of the rhyolite section of the cliff face has been polished to glasslike smoothness, and the underlying tuff has been carved into all manner of fantastic forms. The erosive action has been confined to a section not over 20 feet above the accompanying bench. The sculpturing is probably the result of both wind and water action. The field evidence indicates that water has flowed

along the base of the cliff, accounting for the broad bench occupying that position; also many of the depressions in the cliff face seem to be accounted for only on the basis of water action. Solution pockets are present ranging in size from small depressions up to recesses 2 feet in diameter and 18 inches in depth. Many of the depressions are identical in form with stream-bottom potholes. On the other hand, the waxlike polish exhibited on the face of the harder rhyolite which overlies the tuff appears to be due to the work of wind, as do some of the minor undulations on the face of the tuff.

The bench at the base of the cliff is 100 yards long and 50 feet wide (Fig. 4*b*); upon it rest glacial boulders 4 feet in diameter. The counterpart of this bench can be projected across the valley where it extends into the Gardiner Valley as a spur which is the northwest continuation of the Sheepeater Cliffs (Fig. 1). Below this a succession of recessional benches, composed of coarse gravel and boulders, are traceable along the valley slope and can be projected across to the opposite slope. These benches seem to outline the margins of the shrinking tongue of ice that occupied the valley as it receded eastward.

DISCUSSION

The evidence would seem to indicate that the vast flow of ice during earlier glaciations as well as during the Wisconsin stage, moving northwest in the Yellowstone Valley, was crowded westward and southwestward by the lobes moving into the valley of the Lamar, Slough Creek, and Hellroaring valleys. The character of the drift in the Old Lava Creek Valley, granite boulders from the Beartooth Range, breccias from the Absarokas, and fragments of petrified wood from Specimen Ridge or west of Roosevelt Camp quite definitely indicates the source of the material.

The absence of any sedimentary rocks is to be expected as the thick volcanic deposit to the east probably had not been dissected deeply enough to expose the sedimentary rocks below. Only in Soda Butte Creek and two other very small areas has the limestone been exposed and it is quite probable that this exposure was accomplished by postglacial erosion. It is further evident that the ice from the Gallatin Mountains on the west did not enter the valley from that direction because these mountains are sculptured almost entirely in

sedimentary rocks, and also the train of travertine from Terrace Mountain quite definitely establishes that limit of advance.

At one or more stages in pre-Wisconsin time ice apparently occupied Old Lava Creek Valley from the same source as at the last glaciation. The remnants of the old drift indicate its eastern origin. Whether this was deposited in early or middle Pleistocene time can only be inferred. Alden has suggested:

> In the early part of the glacial epoch, Gardiner River Valley was probably something like 1000 feet less deep than now. It appears fairly certain that Gardiner River Valley had not been deepened sufficiently for the main travertine terraces, on which are the headquarters buildings at Mammoth Hot Springs, to be formed prior to the first stage of glaciation. Possibly it was not that deep before the second stage of glaciation. The pre-glacial bed of Gardiner River may then have been but a few hundred feet below the level of the bottom of the lava capping Mount Everts instead of 1400–1600 feet below that level as now.[4]

The floor of Old Lava Creek Valley is at an elevation 6,600 feet above sea-level and approximately 700 feet below the base of the rhyolite capping Mount Everts. The elevation of the travertine terraces referred to by Alden is 6,238 feet above sea-level. The floor of Gardiner Valley is 5,800 feet above sea-level or 800 feet below Old Lava Creek Valley at the junction of the two valleys. It seems reasonable to assume that erosion of the valley 800 feet in depth would not have been accomplished in less time than that since the Illinoian or possibly the Kansan glaciation.

Apparently Lava Creek was blocked from its eastern exit when the eastern lobe of ice advanced into the valley and forced it to flow along the western margin of the ice lobe into the Gardiner River drainage.

The early ice probably assumed proportions approaching or exceeding the Wisconsin advance. As yet no evidence has been found in the Park at higher levels to indicate this, for, with possibly three exceptions, the highest sections of the central part of the park area were ice-covered in Wisconsin time.

With the recession of the pre-Wisconsin ice fronts, great floods of water must have poured out through the Gardiner, Yellowstone,

[4] W. C. Alden, "Glaciation of Yellowstone National Park and Its Environs," *Ranger Naturalists Manual 16* (1928), p. 61.

and old Lava Creek valleys removing much of the drift, as was the case during the last recession. Evidently the volcanic tuff underlying the rhyolite flow on Mount Everts was in part removed and the cap of rhyolite slumped into the soggy underlying mass. Fragments of the rhyolite became incorporated in this mass as is evidenced there today. Following deglaciation the tuff became cemented as it now appears.

Conditions of a similar nature were apparently reproduced during the advance of the last glacier, the Wisconsin. Ice was crowded westward along the margin of the overfull Yellowstone Valley. A lobe of this flow moved west up the old Lava Creek Valley and a portion of it surmounted Mount Everts. It advanced until it encountered the mass of ice from the west heading in the Gallatin Range. The abrupt termination of drift containing granite boulders from the east and northeast, and drift containing sedimentary rocks from the west, together with the deployment of travertine blocks across the old Lava Creek Valley and the oversteepened west face of Mount Everts with the Gardiner River dissecting the heavy drift deposit at its base, is the evidence for this interpretation.

The mantle of drift from the east and northeast covering Mount Everts and the elevated tracts south and east of Osprey Falls indicate that the thickness of the ice in the Old Lava Creek Valley was at least 900 feet. The presence of drift at the 8,000-foot level on Bunsen Peak, although from another source, suggests the addition of another 400 feet to this figure.

With the gradual melting of the ice, Lava Creek was forced to continue to flow into the Gardiner Valley around the west end of the lobe, clearing a channel through the drift that had accumulated during the recent advance.

The ice in Yellowstone Valley was much thicker than the ice in Old Lava Creek Valley, hence the last to clear; and much drainage would pass along the margin of the ice through Old Lava Creek Valley to Gardiner Valley as the lobe shrank. This would be particularly true as long as ice from the east remained against the eastern spur of Mount Everts blocking most of the northward drainage from the northeastern slope of the mountain. It is probable that the erosion of the tuff beneath the rhyolite cap on Mount Everts and

the broad drainage bench accompanying it are the result of this drainage.

The recessional stages of the last ice are indicated upon the valley sides by a succession of low moraines and terraces (Fig. 2*c*). These moraine deposits have been removed from the valley floor proper and no trace of them is distinguishable less than 30–50 feet above the valley bottom.

The fact that the valley floor east of the col is swept relatively clean of both late and early drift with the exception of a few of the larger boulders indicates that considerable water went through this old channel in the closing stages of ice recession.

The thick drift in the vicinity of Undine Falls may have temporarily acted as a dam until the combined efforts of Lava Creek and the waters held in the old valley by the wall of eastern ice cut it away. It is certain that the drift in the valley near its junction with Gardiner River bears evidence of sorting by running water.

MODIFICATION OF INCISED MEANDERS BY FLOODS

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ABSTRACT

The structurally controlled meanders of Coy Glen were modified in 1935 by a flood of short duration but unusual volume and force. Little change had taken place in the gorge since the original study in 1929, but after the flood it was found that the meander spurs had been truncated and other significant changes had occurred. As slow incision, structually controlled, formed the spurs, their modification in a single flood is definite evidence that such erosion is extraordinary.

INTRODUCTION

In the early spring of 1929 a detailed survey was made of Coy Glen, one of the numerous postglacial gorges of the Finger Lakes region, situated about a mile south of Ithaca, New York. The results of this study were published,[1] but the contour map which formed the basis of the survey was not included as the areal-interpretation diagram was considered sufficient at that time to illustrate the features displayed in this area. A portion of the map made in 1929 is reproduced here (Fig. 1) accompanied by sketches drawn from measurements made in May, 1936. Comparison of these sketches with the map shows that certain marked changes have taken place. Before the unusual flood of 1935 described by von Engeln,[2] little change had taken place in Coy Glen since the original survey.

MAJOR EROSIONAL MODIFICATIONS

The most striking change in configuration of the bed-rock[3] features within the gorge is the truncation of the projecting meander spurs. The amount of material removed and the change in shape are astounding. Some concept of this change may be obtained by comparing the 1929 photograph of spur *F* with the one taken in 1936

[1] W. S. Cole, "Interpretation of Intrenched Meanders," *Jour. Geol.*, Vol. XXXVIII (1930), pp. 423–36.

[2] O. D. von Engeln, "Flood Erosion," *Geol. Soc. Amer. Proc. for 1935*, p. 115 (abst.).

[3] Cole, *op. cit.*, p. 429.

(Fig. 2). Other prominent spurs show like changes in form. This may be noted by comparing spurs *N* and *O* as contoured on the 1929 map with their present configuration as illustrated by the 1936 sketches.

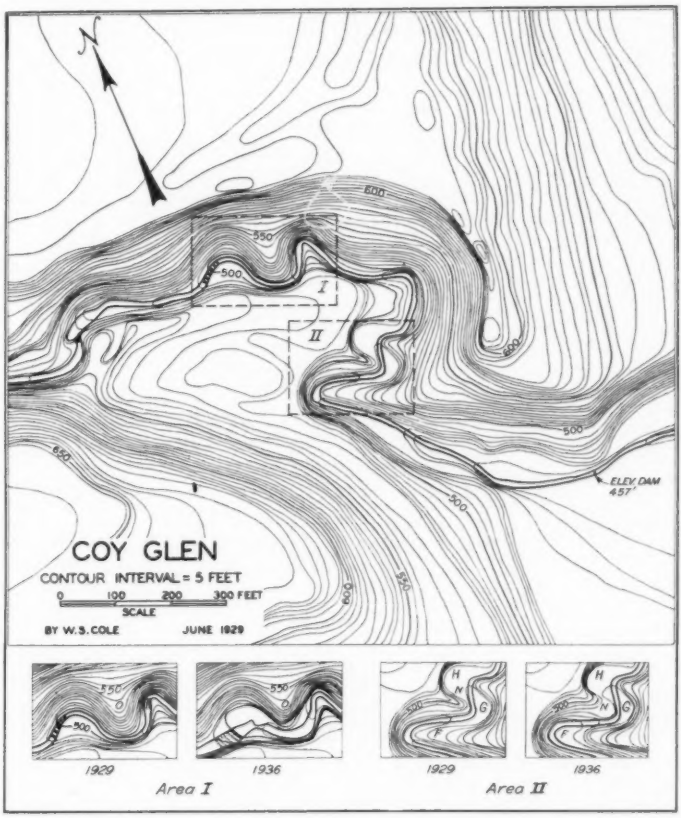

FIG. 1.—Map of a portion of Coy Glen (1929) with sketches showing modification in 1936.

The master joints which control the meander at *K*[4] were formerly concealed in the slip-off slope so that they could be observed only at

[4] *Ibid.*, Fig. 2, p. 425.

K and in the stream bed and gorge wall opposite spur *O*. At the present time they are exposed for their full length across the gorge. Upstream from *K* the more westerly joint forms a vertical wall of about 39 feet in length and some 20 feet in height, thus forming at the present time an east-facing wall in this portion of the meander. Figure 3 illustrates the changes which have occurred, particularly the truncation of the spur opposite the master joints.

FIG. 2.—Meander spur *F* in 1929 (left) and in 1936 (right)

Besides the truncation of spurs, removal of rock occurred in other favorable places. The original photograph of the "sluice-way" falls[5] shows a considerable mass of shale to the right of the two master joints which form the sluice way. Excavation occurred here which removed the greater portion of this shale, producing a marked basin in the area of the former cascading portion of the falls.

AMOUNT OF MATERIAL REMOVED

Approximate estimates of the actual amount of material removed can be given. Spur *F* had material removed over a distance of 60 feet and to a maximum depth of 18 feet at the inner edge of the

[5] *Ibid.*, Fig. 6, p. 431.

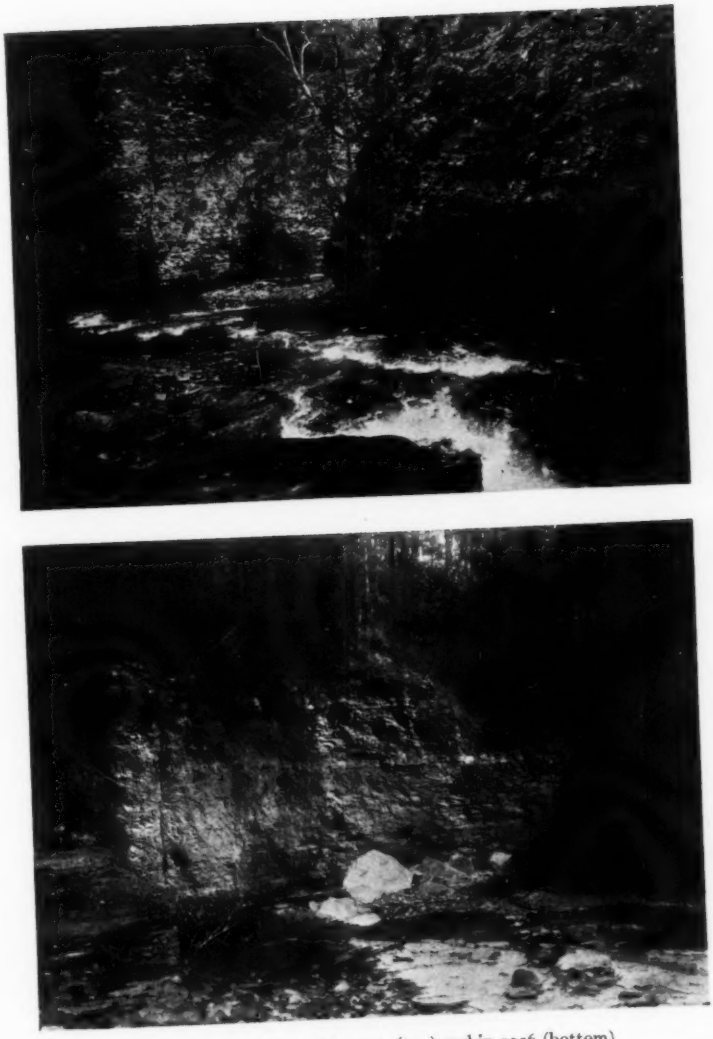

FIG. 3.—Meander at *K* in 1929 (top) and in 1936 (bottom)

removed cone. The estimated loss from the spur would be approximately 50 cubic yards. Spur *N* was cut back a distance of approximately 50 feet and to a depth of about 15 feet at the inner edge of the removed area. Average depth of removal throughout the entire length would approach 7 feet. The estimated loss from this spur is about 125 cubic yards. The spur at *O* had material removed from a rude triangular area. The average depth of removal over this area is about 5 feet with an approximate total loss of 40 cubic yards. The total loss from the three measured areas is approximately 215 cubic yards. As this figure is for three very limited areas, the total removal during these floods would be a very considerable amount.

The spur opposite meander *H* had been truncated before 1929 because of the accumulation of gravel in the meander bend which forced the stream into a course which gradually removed the spur while the meander bend was protected. This spur suffered little change in the 1935 flood.

In general, widening of the gorge was accomplished by the removal of projecting spurs. It is notable, however, that the meander bends were not markedly increased in size. The distance across the gorge at *K* along the more westerly joint in 1929 was approximately 110 feet. This is practically the same distance as obtained in the present measurement. The meanders at *G* and *H* also have the same proportions. These meanders, however, were protected by gravel bars, while the meander at *K* was entirely unprotected.

MAJOR DEPOSITIONAL CHANGES

In 1929 the occurrence of gravel bars within the meanders had diverted the stream so that it was not enlarging either of the meanders *H* and *G*[6] but, instead, was truncating the noses of the spurs opposite these swings. At this time there was indication that the stream might resume its former course through this area because a channel had opened during the unusually high water of that spring across the gravel bar. With this channel available it was expected that the full current would be very shortly flowing in its old course again. In 1934 the condition described above had changed but slightly. The logs which had caused the original deposition of the gravel within the meander were still in place.

[6] *Ibid.*, Fig. 2, p. 425.

The present course of the stream is nearly identical to that of 1929 and 1934 with the exception that the minor channel across the bar is missing. This is unexpected as it was considered that a major flood would remove the gravel piled in back of the log jam so that the stream could resume its former channel. The logs were removed by the recent floods, but great masses of gravel and sandstone blocks occupy the area of the old gravel bars. The total mass of debris is greater than before.

During the first rush of water the meander bends were occupied and, without doubt, the accumulated debris was swept out. Following this, these bends apparently became pockets of deposition into which debris was hurled while the channel around, and partially over the tops of, the meander spurs was maintained. In fact, it is notable that most of the cutting occurred on the spurs rather than on the outer edge of the meander bends.

GEOLOGIC ACTIVITY OF FLOODS

Under normal conditions of weathering and erosion the truncation of the meander spurs would not have been accomplished in many years. Because of the unusual volume of water the entire width of the gorge was occupied during this flood. Thus the deflectional effect of the meander bends was lost. The main current, instead of following the outer edges of the bends, cut diagonally across these areas so that the entire force of the current was directed at the spur noses rather than toward the cut banks opposite the spurs.

The removal of the spurs was greatly facilitated by the weathered condition of the shale comprising the spurs. The bed rock of the area is dominantly shale with small sandstone members. As these shale beds are cut by numerous joints and have closely spaced bedding planes as well, the agents of weathering had easy access. The depth and extent of this weathering can be observed by iron stains on the facets now exposed by the removal of the major portions of the spurs.

PRESENT IDEAS COMPARED TO THOSE PREVIOUSLY HELD

The main thesis of the development of Coy Glen as presented in the original article is completely in accord with the continued development of the gorge. The incipient meander at M^7 had developed

[7] *Ibid.*, Fig. 2, p.425.

considerably, especially under the acceleration of erosion during the floods. Other structurally controlled meanders are developing at points previously considered suitable for their development, especially upstream from the area studied in detail.

The fact that the spurs might be truncated and largely removed by floods was not appreciated at the time of the original investigation. The configuration of the spurs indicated that they had been formed during a long period of slow incision under the guidance of structural control. The extreme modification of these spurs in a single flood is evidence that such erosion is extraordinary and presents a striking contrast to the concept previously held that they would be gradually removed by weathering, supplemented by minor stream shifts with impingement on the spurs.

The partial removal of the spurs gives the gorge a more open and mature appearance than it formerly had. The master joints, however, are still the dominant control, and the distinctive meander pattern is still the outstanding feature of the gorge. Even such accidents as befell the gorge in 1935 will not prevent the development of structurally controlled meanders although their incipient form may be modified.

CONTRIBUTIONS TO PANAMA GEOLOGY

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ABSTRACT

The central highlands of Los Santos province, South Panama, a complex of acid to basic igneous rocks, together with tilted Eocene beds, was the scene of volcanic and mountain-building activity, probably about the close of the Oligocene. Cessation of volcanism may not have occurred until Pliocene or Pleistocene time.

INTRODUCTORY

The province of Los Santos forms the eastern part of the Azuero peninsula, which juts far southward into the Pacific from the west-central part of Panama. The province, fairly populous, is connected with Panama City, 95 miles toward the north-northeast, by weekly steamer. The writer visited the area in 1918. More recently R. A. Terry and A. A. Olssen made a 25-mile reconnaissance traverse from Las Tablas over the mountains to Tonosi, collecting some rock specimens on the way. Thin sections of these were studied by John J. Ronan under the direction of the writer.

PHYSIOGRAPHY

Las Tablas, situated inland a few miles from the eastern coast of Los Santos province, is among very low rolling hills, remnants of a coastal plain. About 7 miles from there the traverse, south-south-westerly in direction, led to the edge of a low but rugged mountain area. This—some 15 miles long by 8 miles wide—forms the central highlands of Los Santos. Irregular ridges separated by crooked steep valleys and swift-streamed canyons, with some steep peaks reaching up to 2,500 feet, characterize the area as one of very youthful topography. A few peaks, roughly conical in outline, stand many hundreds of feet above the surrounding hills. Their dark-gray rocky upper slopes, well shown on Cerro Quema, indicate a relatively recent volcanic origin. In general, these highlands are largely covered with tropical forest, except on the steeper slopes. This, together with obscuring soil and the great variety of volcanic rocks,

is the chief reason why the hurried traverse, through a very sparsely settled area, was able to elucidate only the broader features of the complex geologic history of the region. From the highlands the traverse descended into the wide flattish Tonosi Valley, locally

FIG. 1.—Los Santos province, part of the Azuero peninsula, southern Panama. It is situated approximately a hundred miles south-southwesterly from Panama City.

floored with Eocene limestone. On this rich soil there is considerable cultivation. The valley is drained by the small Tonosi and the much larger Guere rivers and contains the little town of Tonosi, where the traverse ended.

SEDIMENTARY ROCKS

Sedimentary rocks occur at both ends of the traverse dipping away from the central igneous complex. The limestones of the Tonosi area are apparently a part of the Eocene series (sandstones, shales, and limestones) which outcrops along the shore in the vicinity of Guanico Point, about 5 miles southerly from Tonosi. The sandstones and shales near Las Tablas dip away from the central igneous mass so steeply that they are almost vertical.

LIMESTONES

The Eocene limestone sample studied contained 9.97 per cent of insoluble matter. It is very fine-grained, breaks with a conchoidal fracture, and is dull slaty brown in appearance. Under the microscope fine, parallel, almost straight, lines indicated bedding laminae. Along these laminae are some very minute grains of dark to opaque mineral matter, probably fragments of ferromagnesian minerals, somewhat altered. Many of these are enclosed in very thin concentric films of calcite. The appearance of this particular specimen under the microscope suggests chemical precipitation. Fossils from these rocks were studied by A. A. Olsson (Specimen No. 3, Traverse Station 363).

SHALES

A limy shale, of Eocene age, apparently lies under the limestone. Specimen 7, Traverse Station 441, contains 14.53 per cent of soluble matter, is fine-grained, breaks with a conchoidal fracture, and is slaty gray to greenish in color. Under the microscope very fine quartz grains were observed, together with a considerable amount of very minute dark opaque grains, apparently ferromagnesian fragments, now largely altered to chlorite, the latter giving the rock its slightly greenish tint. Fine bedding laminae were observed.

Specimen No. 9, Traverse Station 552, of the same general limy shale, contains 32.83 per cent of soluble matter. Under the microscope fragments of plagioclase, some orthoclase, and weathered ferromagnesian particles were observed. A few foraminifera were noted. The specimen contains several very small veinlets of calcite, is dark gray in color, and fairly fine-grained.

Specimen No. 5, Traverse Station 298 (?), is that of a nonlimy

meta-shale, considerably altered and somewhat mineralized. Minute crystals of pyrite are disseminated throughout the dark-green specimen. Under the microscope colorless to pale yellow epidote and considerable chlorite were noted, also tiny veinlets of secondary quartz. There is no evidence available to the writer as to the age of this shale. It may be early Eocene or Cretaceous, having been altered by younger intrusive rocks.

IGNEOUS ROCKS

The highland area crossed by the traverse is mostly a complex of acid to basic igneous rocks, the relations of which have not yet been worked out. The youngest of the intrusive rocks is post–early Eocene because the Eocene beds dip away from the igneous area.

TABLE 1

	Per Cent
Orthoclase	14.12
Albite	25.86
Quartz	37.50
Hornblende	0.72
Chlorite	0.25
Calcite	0.20
Epidote	18.35
Magnetite	2.50
Secondary white mica	0.50
Total	100.00

Cerro Quema, rising nearly a thousand feet above its surrounding hills and bare of vegetation in its upper part, apparently was a volcano in very late Tertiary or in Pleistocene time. Its dark-looking rocks suggest andesitic pyroclastics. The age relations of the granitic and the basic rocks of the area are not known, though the writer is inclined to the view that the granite is the older.

GRANITES

Specimen No. 8, Station 237, represents a light-colored, fine-grained, sodic granite, which weathers rather pinkish. Under the microscope the percentages of minerals were determined with the aid of a Leitz integrator for measuring mineral areas in thin sections (Table 1).

The orthoclase and albite, present in both large and small crystals, have been considerably altered to epidote, calcite, and sericite. Some micrographic intergrowths of quartz and orthoclase were noted. Magnetite appears as a primary mineral and is also secondary, associated with chlorite, from the alteration of hornblende. The alteration of this granite may be due to later igneous intrusions.

GABBRO

Specimen No. 1, Station 187, represents a quartz-gabbro. A micrometric analysis of it gives the mineral composition as shown in Table 2.

TABLE 2

	Per Cent
Labradorite	51.02
Hornblende	7.92
Quartz	27.93
Magnetite	3.20
Sphène	1.02
Calcite	0.73
Chlorite	5.43
Epidote	2.50
Secondary white mica	0.25
Total	100.00

Labradorite feldspar occurs in well-shaped crystals having a marked zonary structure and twinned on both the Carlsbad and the albite laws. It is partly altered to epidote, calcite, and white mica. The hornblende, pleochroic from greenish-brown to green, has been altered, in some cases entirely, to chlorite. Quartz occurs, as in granite, interstitial between the earlier formed crystals. The magnetite is present as regular octahedrons and in some irregular grains. Much of the sphene shows rhombic cross sections. An alteration characteristic of this rock is fine-grained specular hematite disseminated along cracks in the rock and penetrating the solid rock on each side of the crack to a depth of about $\frac{1}{4}$ mm.

The relation of this quartz-gabbro to the granite is not known. One would suspect, however, that they are not far separated in age and that younger intrusives altered both rocks.

ANDESITES

Specimen No. 6, Station 178, is from a hornblende andesite. The composition of the rock is shown in Table 3.

The andesine crystals show rounded corners with twin lamellae not distinctly outlined. Alteration products, not resolvable, give it a very turbid appearance. The hornblende, pleochroic from yellow to green, is present in crystals which range in size from microscopic to a centimeter in width. Many show marked resorption borders, and most of them have been partly or entirely altered to chlorite.

TABLE 3

Phenocrysts	Per Cent
Hornblende	16.0
Andesine	26.5
Quartz	3.8
Magnetite	1.8
Sphene	1.0
Groundmass	50.9
Total	100.0

Quartz occurs in well-rounded grains, and euhedral crystals of magnetite are present. The fine-grained grayish groundmass of the rock contains crystals of andesine and hornblende of recognizable size.

PYROCLASTICS

Specimen No. 2, Station 118, is an andesite breccia. The andesine crystals present are partly altered to calcite, white mica, and epidote, so that the twin lamellae appear discontinuous. The primary augite, present in some instances, has the typical cleavage traces of diallage and occurs in relatively large fragments which enclose small bodies of other minerals, giving a poikilitic texture. It has been altered to a considerable extent to chlorite. Quartz is present in irregular fragments. Magnetite also is fairly common.

Small angular fragments containing one or more of the minerals augite, epidote, and chlorite are scattered throughout the rock and indicate its pyroclastic nature. The rock is dark green in color, due to the great amount of chlorite present, with large dark fragments of

other rock, and also yellow grains of epidote. The groundmass is andesitic and very fine in grain.

Specimen No. 4, Station 162, closely resembles No. 2 except for its finer texture. It is also an andesite breccia but with its brecciated fragments less conspicuous. This breccia may have been formed from showers of hot rock fragments, the temperature causing them to blend together fairly well. The hand specimen is dark green in color with small, angular dark and light patches. Under the microscope the andesine feldspar appears highly altered to epidote and white mica. Some of the almost colorless augite is completely altered to chlorite. Accessory magnetite is also present.

LAVA FLOWS

The last two specimens studied—No. 10, Station (?), and No. 11, Station 56—are from amygdaloidal lava flows. Specimen No. 10 is fine-grained and dull gray in color with light-colored amygdules of varying sizes. Under the microscope the vesicles are seen to be filled with spherulitic chalcedony and in a few cases with chlorite in radiating aggregates. The amygdules vary in size up to 2 cm. or more in diameter. Augite, orthoclase, plagioclase, and magnetite are present in small crystals in the groundmass. The feldspars, in laths, appear to be arranged in more or less radiating aggregates.

Specimen No. 11 is dark gray in color, fairly fine-grained, and has spots of zeolites scattered through it. Under the microscope the vesicles are seen to be filled with zeolites which are of three kinds: analcime, heulandite, and stilbite. In some instances all three are associated. The analcime, colorless to light yellow, occurs in very fine sheaflike aggregates. Stilbite and heulandite usually occur associated with each other, the former in colorless sheaflike aggregates and the latter in small crystals. The labradorite feldspar present is almost entirely altered to calcite, epidote, and white mica. Chlorite, from a ferromagnesian mineral, and magnetite are also present. The rock is classified as a basaltic lava flow.

CONCLUSIONS

Owing to the fact that beds of known Eocene age are tilted away from the central mountain area, a post–early Eocene mountain-

building period is indicated. However, basic dikes and volcanic necks filled with andesite-breccia cut Oligocene beds where the Panama Canal passes through the continental divide. These beds also contain a lava flow and are cut by some basalt, but no acid igneous rocks are exposed there. At the north end of the Canal extensive Miocene beds occur near Gatun, but these have not been cut by igneous rocks, so far as the writer could find. These conditions, only a hundred miles northwest of Los Santos province, suggest that its chief mountain-building also occurred at about the close of Oligocene time. However, such extinct volcanic cones as Cerro Quema, little modified by erosion, would seem to indicate that volcanic activity did not cease in the area until Pliocene or even Pleistocene time.

STUDIES FOR STUDENTS

SHIFTING OROGENIC BELTS OF THE SOUTHERN CANADIAN SHIELD

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ABSTRACT

An attempt is made to utilize what has been learned from the behavior of geosynclines and the manifestations of typical mountain folding with batholithic intrusions, in later geologic times, to better our understanding of the pre-Cambrian history of the Canadian Shield. An attempt is made to locate roughly the sites of the Laurentian, Algoman, and Killarnean foldings and granitic intrusions. Erosion of each of these mountain systems should have supplied sediments in great volume to the neighboring lowland tracts until the highlands themselves eventually were brought low. Where were the lowland tracts receiving abundant sediments from each of these mountain systems?

The Laurentian Mountains seem to have comprised a southern system, south of Lake Superior, and apparently also other ranges far north of Lake Superior. Rock material from their denudation was spread upon a broad sinking belt to the north of the southern system and in intermontane belts still farther north, producing the Timiskaming–Knife Lake sedimentary system. Later the Algoman folded ranges arose where the weak Timiskaming sediments had accumulated in great thickness. From the northerly Algoman Mountains, sediments were carried southward into the Penokean trough, south of present Lake Superior, in a part of the old Laurentian area. Here accumulated the Lower Huronian formations, and later the Middle and Upper Huronian systems were spread over somewhat wider areas. Subsequently the Killarney folding and igneous injection occurred in this southern belt of extensive Huronian deposition.

INTRODUCTION

While considering different interpretations of the pre-Cambrian of the Canadian Shield with a class of advanced students early in 1935, there flashed the idea that one might apply the recognized principles of later orogenic behavior to these pre-Cambrian problems. With the geologic map of North America before us, we attempted to locate in a rough, general way the positions of the successive belts of mountain folding in the southern part of the Canadian Shield and with them the neighboring tracts that should have received thick sedimentary deposits from the denudation of each of these mountainous strips. The outcome was surprising. In this hasty improvisation the larger picture of the pre-Cambrian history of this great province seemed to unfold in an apparently consistent

synthesis. The better-known facts appeared to fit together logically, as they should in a correct interpretation.

For two years nothing further was done to test the validity of these generalizations by checking with the available maps and reports covering the many individual districts involved. However, a few months after this classroom improvisation there appeared the *U. S. Geological Survey Professional Paper 184*, "Pre-Cambrian Rocks of the Lake Superior Region," by C. K. Leith, Richard J. Lund, and Andrew Leith, which brought up to date in admirable summary form the geology of this large area and supplied an excellent revised map. In this long-awaited publication some of the differences in the distribution of the Laurentian, Algoman, and Killarney granites were brought out clearly in the text and made quite apparent on the map. The adopted correlations also appear to mark a long step toward a satisfactory agreement on many debated questions. They have accordingly removed part of the excuse for the present discussion. But in the belief that it may still prove helpful to students, confronted with the perplexing conflict of views of pre-Cambrian correlation encountered in the literature, to approach the whole problem from the quite different point of view of orogenic behavior and typical mountain history, the present visualization of the ancient orography of the Lake Superior–Ontario–Quebec region is offered for whatever it may be worth.

TYPICAL OROGENIC BEHAVIOR

Let us apply, in uniformitarian fashion, some of the principles learned from the study of later mountain systems. It is tacitly assumed, and confidently believed, that the principles of uniformitarianism are as properly applied to the development of the very old rock formations and structures of the earth's surface as to the development of those of any later time. Let us get in mind first the evolution and history of a typical mountain system. A familiar example is the Appalachian system. Through most of Paleozoic time the land of Appalachia fringed the United States on the east. It was an orogenic belt which was folded at several different times, was reduced by denudation between episodes of diastrophic rejuvenation, and constituted a long-lived major feature of Paleozoic North

America. Adjoining on the west was the long Appalachian geosyncline receiving many thousands of feet of sediments from the bordering ranges of Appalachia during a large part of the Paleozoic. In the geosyncline the Paleozoic stratigraphic record was a very full one, though interrupted by many unconformities. Late in the Paleozoic came the Appalachian revolution. The tens of thousands of feet of weak Paleozoic sediments in the geosyncline were strongly folded, and folding affected also the stronger rocks farther east. In the heart of the folded belt large batholiths of granitic rock were intruded, some of which are now exposed in the present Piedmont strip. Extensive denudation followed the orogenic revolution. Subsequently, to the east of the present Appalachian Mountains, a long trough of downfaulting developed, and into this the mountains which had risen from the Paleozoic geosyncline farther west shed many thousand feet of Triassic sediments. In this history we see a shifting of the sites of extensive sedimentation in response to shifting manifestations of diastrophism.

Our western mountains, various European chains, and mountain systems in many other parts of the world exhibit likewise, in varying degree, a shifting of the belts of strong orogeny and of the related areas of exceptionally abundant sedimentation. This is expressed in the geosynclinal theory of the origin of folded mountain ranges.

For one more example let us consider Tibet and northern India. Late Paleozoic folding developed mountain ranges in Tibet. To the south, in the Mesozoic, were the Straits of Tethys featured in some of the writings of Eduard Suess. For ages detritus from the Tibetan mountains poured into this long geosyncline. In the Tertiary the geosyncline sedimentary belt was folded and uplifted into the present lofty Himalayas. Since then vigorous streams have transported vast amounts of rock material from the mountains to the bordering plains of northern India. An east-west strip of these plains, now occupied in part by the Indus and Ganges rivers immediately south of the Himalayas, has been sinking in geosynclinal fashion and has already received several thousand feet of late Tertiary and later sediments. The process presumably will continue, and at some distant date in the future when strong compressive stresses again develop in this portion of the globe, we may expect this fringing belt of

thick Indo-Gangetic sediments (Siwalik and later formations) to be folded into a new mountain range south of the present Himalayas. Here is a southward migration of the successive generations of mountains and the belts of thick sedimentary accumulation apparently still in progress.

Another characteristic behavior of strong mountain folding is that, during a late stage of the orogenic revolution, large volumes of acidic magma are injected into the heart of the folded zone. By the time the magma is injected, deposition of the youngest system of rocks which is involved in the main folding has ceased in the folded areas at least. Instead, erosion is now in full sway, and eventually the folded strata overlying the massive intrusions may be removed from large areas exposing parts of the granitic batholiths in the medial portion of the former mountain system. A considerable length of time has been involved in the reduction of the mountains and exposure of the granite masses. This is represented today by a major unconformity. Subsequently deposition may again plaster sediments over the beveled structures and batholiths of the old ranges, giving rise to a younger superjacent rock system.

The pre-Cambrian rock systems today are separated and classified by the great unconformities which represent long intervals of time between periods of prolonged sedimentation. Equally available for separation and classification of these systems are the extensive granitic intrusions, for, in each case, they represent a part of the same interval of time as the corresponding unconformity. Consequently separating rock systems by unconformities, or classifying them by granites, really leads to the same results. Actually it may be easier to work with the granites than with the unconformities. Hence their great importance and utility in pre-Cambrian classification and correlation.

With these principles and modes of behavior clearly in mind let us turn to the Canadian Shield. In the southern portion of this large area at least three pronounced pre-Cambrian orogenic revolutions with accompanying granitic intrusions have been recognized. These were the Laurentian, Algoman, and Killarnean orogenies. Naturally they have been taken to mark off major divisions of the pre-Cambrian history of the region.

THE LAURENTIAN GRANITES

Where were the Laurentian Mountains? From the strongly folded structures of the pre-Laurentian rocks and the less deformed overlying rocks, their location eventually will be determined, so far as this is possible. In the meantime we may use the Laurentian granites to locate the core of this ancient mountain system, recognizing, however, that the folded belt was probably somewhat wider than the zone of intruded granites. Where have the Laurentian granites been recognized?

On the new map of the Lake Superior region[1] Laurentian granite is shown prominently on the south side of the lake. In both the Marquette and Penokee-Gogebic iron districts it is much in evidence, extending southward from Lake Superior, near Marquette, to the Menominee River, and from the Gogebic district for an unknown distance toward central Wisconsin. Granite crops out, or underlies the drift, over a large part of northern Wisconsin, but much of this is mapped as undifferentiated pre-Cambrian granite; in a few places as far south as Wisconsin Rapids it is designated Laurentian and in others doubtfully Killarnean. None of it is called Algoman; in fact, no Algoman granite is represented anywhere south of Lake Superior. The map's structure section from Thunder Bay, Ontario, across Isle Royale, Keweenaw Point, and the Marquette and Menominee districts to Green Bay show Algoman granite on the north side of Lake Superior but on the south side enormous masses of Laurentian granite intruded by smaller batholiths of Killarney granite.

On the north side of Lake Superior the granite occurrences are thus quite different from those on the south side. Some narrow strips of Killarney granite are associated with the Duluth gabbro not far from the lake, but the large areas of granite north of the gabbro mass and north of the Mesabi iron belt are mapped as Algoman as far as the international boundary, with the exception of the small Saganaga intrusion at the boundary line. Early workers considered most of this granite to be of Laurentian age, but now only the Saganaga granite, appearing in a small oval area of about 200 square miles partly in Minnesota and partly in Ontario, is thought to be older

[1] C. K. Leith, R. J. Lund, and A. Leith, *U.S. Geol. Surv. Prof. Paper 184* (1935).

than the Ogishke–Knife Lake sediments and hence called Laurentian.[2]

Farther north in Ontario, Merritt has mapped considerable Laurentian in a strip along the Canadian National Railway, but regards this classification as tentative since these granites are only known to cut the Keewatin.[3] Isolated patches of very old granite have been found in other areas considerably farther north. Whether they all belong to one period of intrusion, or to several different times of intrusion, is not yet known. Independently of these exposures of granite, however, Pettijohn has recognized the necessity for postulating strongly elevated, east-west trending granitic ranges in western Ontario, to supply the boulders and cobbles found in great abundance in belts of extremely thick pre-Huronian conglomerates, which he thinks must have accumulated in basins or troughs between the ranges.[4]

On the east side of Lake Superior two granites have been recognized in the Michipicoten iron district, though no attempt has been made to map them separately. "It is supposed, however, that the pre-Dorean granite is not extensive and that the main body of granite in the western and northern part of the map area is younger than the Dorean and post-Dorean divisions of the schist complex."[5] From Michipicoten Harbor south to Batchawana Bay the granites are all labeled "Algoman or Laurentian" on the 1935 map of the Lake Superior region. It is to be noted that this strip lies between the area of dominantly Algoman granite to the north and that of dominantly Laurentian granite to the south. Just south of Batchawana Bay and toward Sault Ste Marie small intrusions of Killarney granite appear.

It is stated in *Professional Paper 184* that the Laurentian granite is widespread throughout the region, the Algoman is confined largely

[2] Frank F. Grout, "The Saganaga Granite of Minnesota-Ontario," *Jour. Geol.*, Vol. XXXVII (1929), pp. 562–91.

[3] Phillip L. Merritt, "Seine-Coutchiching Problem," *Geol. Soc. Amer. Bull. 45* (1934), pp. 333–74.

[4] F. J. Pettijohn, oral communication.

[5] W. H. Collins, T. T. Quirke, and Ellis Thomson, "Michipicoten Iron Ranges," *Geol. Surv. Can. Mem. 147* (1926), p. 36.

to the north shore of Lake Superior, and the Killarney granite occurs mostly on the south shore.[6] Nevertheless, the Laurentian, if present identifications are correct, appears much more abundantly on the south shore than on the north shore. From this one may conclude that the heart of an old Laurentian mountain system lay south of the present site of Lake Superior. Lesser outlying ranges may, of course, have flanked it on the north. Just as the better-understood Laramide orogeny developed isolated outlying ranges like the Bighorns, Black Hills, etc., at considerable distances from the main Rocky Mountain system, so the old Laurentian orogeny likewise may have produced lesser isolated ranges to the north of the main folded system.

The existence of large, more or less paralleling ranges still farther removed to the north, and separated from the southern Laurentians by a geosynclinal tract in which the Knife Lake and Seine River sediments accumulated, has been inferred by Pettijohn.[7] Apparently there were several of these very old northerly ranges with intermontane belts of coarse, clastic deposition. That they were all, or in part, formed during the same period of orogeny as the Laurentians south of Lake Superior is, however, by no means certain.[8] Some may be older (as suggested by the Rice Lake granite for which radioactive age determinations give an age of 1,800 million years) and some may be younger.

From the region of Sault Ste Marie the northern boundary of the southerly Laurentian granite area seems to continue eastward south of Sudbury and Lake Nipissing to the Ottawa River where it presumably curves to the northeast, as indicated by the regional structural trends. The southern borders unfortunately are now obscured by later Paleozoic formations. Basing our view upon the information available at present, we may therefore visualize these ancient Laurentian folded ranges as extending eastward across northern Wisconsin, Michigan, the Lake Huron–Georgian Bay area, south-

[6] Leith, Lund, and Leith, *op. cit.*, p. 1.

[7] F. J. Pettijohn, "Early Pre-Cambrian Geology and Correlation Problems of the Northern Subprovince of the Lake Superior Region," *Geol. Soc. Amer. Bull. 48* (1937) p. 192.

[8] F. J. Pettijohn, oral communication.

eastern Ontario to the original Laurentian area of Quebec, and thence northeastward toward Labrador.

But we ought to be prepared, however, for another possibility. As already noted, the tendency of later studies has been to show that various granites, once thought to be Laurentian, are in reality of later date. Should future investigations take from the Laurentian more and more of the granites south of Lake Superior and from there eastward, the southerly Laurentian Mountain ranges, inferred from present mapping of the granites, would diminish correspondingly in importance. In that event relatively more of the succeeding Timiskaming sediments might have to be attributed to the denudation of more northerly Laurentian ranges, such as those postulated by Pettijohn.

TIMISKAMING SEDIMENTARY BELT

Erosion of the Laurentian Mountain belt inevitably gave rise to sedimentary deposition in the neighboring lowland areas. Such deposition of detritus may have been either north or south of the highland strip or on both sides of it. Studies of later mountains in various parts of the world indicate that, as a rule, deposition has been much more abundant on one side of a mountain system than on the other. Very commonly there has been a geanticlinal-geosynclinal couple; the elevated mountain belt has been paralleled on one side by a sinking geosynclinal trough in which sediments from the mountains have accumulated to exceptional thicknesses. We should look for such behavior alongside the Laurentians.

The southern environs of the ancient Laurentians unfortunately are not now exposed to study owing to the covering of Paleozoic formations. On the north side of this southerly range, however, are widespread outcrops of the thick Timiskaming sedimentary rock system. Here is the great mass of neighboring post-Laurentian sediment which we are seeking. It is very abundant in the Lake Timiskaming region. Around Sudbury no Laurentian granite has been found, suggesting that this area also lay outside the main Laurentian disturbed belt. Of like import is the fact that the Sudbury series, some 7,000 feet thick, rests on the Keewatin without angular unconformity.[9] Northwest of Lake Superior is the Knife Lake series

[9] W. H. Collins, "Sudbury Series," *Geol. Soc. Amer. Bull.* 47 (1936), pp. 1675–89.

and its equivalents. Their material came from the Laurentians and they cover large areas. The many thousand feet of Timiskaming sediments in many places indicate pronounced sinking of the Timiskaming sedimentary belt which we may consider to have been geosynclinal in nature. As the history of younger and better-known folded mountain ranges has commonly been that, after the early phase of strong folding and ensuing extensive reduction by erosion, vertical uplifts have later rejuvenated the orogenic belt, we may well suspect that a succession of later uplifts likewise restored the Laurentian highlands while the adjoining Timiskaming belt was sinking and receiving its heavy burden of sediment.

One of the chief difficulties in the way of satisfactory correlation of the pre-Cambrian of the Canadian Shield has been the fact that the Lower Huronian rests directly on Laurentian or pre-Laurentian rocks on the south side of Lake Superior, as do the Knife Lake series on the northwest side of Lake Superior and the Sudbury and Timiskaming much farther east. It was suggested long ago that these post-Laurentian sedimentary formations are correlatives since they seem to occupy comparable positions in the stratigraphic succession. But, on the other hand, the Lower Huronian south of Lake Superior correlates well with the Bruce series farther east in Ontario which is younger than the Timiskaming, which is separated from the Huronian rocks by a great unconformity. In consequence there has been a perplexing difference of opinion between the geologists working on the south side of Lake Superior and those studying eastern Ontario. The former find the greatest break in the pre-Cambrian between the Laurentian (or pre-Laurentian) and the Lower Huronian; the latter, while they have recognized a great break between the Laurentian (or pre-Laurentian) and the Timiskaming, claim that there is a still greater break between the Timiskaming and the Bruce.

What does our method of study seem to indicate? The Laurentian granites, strongly developed south of Lake Superior, extend right to the shore of the lake near Marquette. These granitic batholiths, intruded presumably in the heart of the Laurentian Mountains, would say to us that the Laurentian Mountain belt occupied the whole present south side of Lake Superior. During Timiskaming time the south side of Lake Superior was an elevated tract, re-

juvenated occasionally as it was lowered by denudation, thus supplying abundant sediments (Knife Lake series, etc.) to the neighboring trough lying farther north, but itself undergoing erosion instead of deposition. Thus the Timiskaming rock system is not found on the south shore of Lake Superior. Timiskaming time is represented there only by an unconformity. Natural it is that this unconformity should be the most pronounced in the pre-Cambrian of that region. The Laurentian revolution and Laurentian intrusives had given great complexity to the older rocks of the area, and the time involved in the unconformity was from the Laurentian revolution, through the long ages of Timiskaming deposition and the subsequent Algoman revolution elsewhere, to the spreading of much later Lower Huronian sediments upon the peneplaned surface of the basement complex. Taking into account these bits of the history, learned from areas outside the belt of Laurentian folding, removes some of the troublesome difficulties that stood in the way of satisfactory correlation. The break between the Timiskaming and the Bruce will be considered in its proper place in the historical sequence.

THE ALGOMAN GRANITES

Let us recall two time-honored principles: (1) the principle of posthumous folding of Suess, who noted that later foldings have in many places followed along nearly the same lines as earlier foldings; and (2) the geosynclinal theory of the origin of mountain ranges, according to which the strongly folded mountain systems arise in belts where previously weak sediments have accumulated to more than average thicknesses. Recurring orogenic episodes separated by moderate intervals of time commonly affect essentially the same belt, and what we call the Laurentian revolution may have comprised more than one paroxysm of strong folding. But after sufficiently longer stretches of time, during which there has been extensive filling of an adjacent geosyncline, the later folding is likely to be most conspicuous in the neighboring geosynclinal belt. Accordingly we may expect to find that the next great mountain system, with accompanying batholithic intrusions, arose chiefly from the Timiskaming belt of exceptional sedimentation north of the older Lauren-

tian Mountains, though it might also have included a portion of the older mountain area.

The succeeding revolution was the Algoman. Where are the Algoman granites which were intruded in the midst of the Algoman Mountains? They are found prevailingly north of the southerly belt of Laurentian granites, intruding the folded Timiskaming beds, although the possibility must be recognized, of course, that there may also be some granites of Algoman age in the Laurentian area south of the Timiskaming border where, in the absence of Timiskaming beds to date them, they have not been differentiated from the Laurentian granite. Northwest of Lake Superior the Giants' Range and Vermilion granites of Minnesota are regarded as Algoman. Farther north in Ontario are large strips of Algoman granite as far at least as the Savant Lake area and Lake Windegokan characteristically intruding belts of very thick pre-Algoman sedimentary rocks. East of Lake Superior Algoman granites occur in the Michipicoten and Mississagi districts. They are absent at Sudbury whose position apparently lay south, though not far south, of the main Algoman ranges, but they are very conspicuous to the north and northeast in the Porcupine, Kirkland Lake, Lake Timiskaming, and other areas. We therefore conclude that the great Algoman Mountain system occupied a broad zone from at least the middle of present Lake Superior on the south more than half-way to James Bay on the north. Its western and eastern extremities are not considered here, but the portion which is best known reached from northwestern Minnesota and Manitoba far eastward into Quebec. It was a northerly mountain system.

THE THREE HURONIAN SYSTEMS AND THE KEWEENAWAN

Erosion of the Algoman ranges in their turn, like that of the earlier Laurentians, shed sediments onto the neighboring lower tracts. On the south shore of present Lake Superior and the north shore of Lake Huron, which are nearly in east-west alinement not far south of the folded Algoman Mountain belt, these formed the Lower Huronian–Bruce sedimentary system. Cooke's instructive map for the Lower Huronian period shows the only known area of Lower Huronian deposition as a narrow trough (Penokean) extend-

ing from south-central Minnesota across northern Wisconsin, northern peninsula of Michigan, north shore of Lake Huron, and eastward practically to the Ottawa River.[10] This Lower Huronian trough thus lay just to the south of, and parallel to, the Algoman Mountain system. Its western portion developed within the old Laurentian Mountain area which had been peneplaned during the long Timiskaming period; its eastern portion was partly on the north flank of the Laurentian area and partly (as at Sudbury) in the less-folded area between the Laurentian and Algoman folded belts where neither Laurentian or Algoman granites are found. In this trough the Lower Huronian sediments accumulated.

Van Hise and Leith recognized long ago, as already noted, that the greatest break within the pre-Cambrian of the Lake Superior region was between the Keewatin-Laurentian complex and the Lower Huronian on the present south shore, and this accordingly was adopted by the United States Geological Survey as the division between the Archean and the Algonkian (Archeozoic and Proterozoic). Why was this break so prominent? Because this was the area of the Laurentian Mountains of strongly deformed rocks which, during the succeeding Timiskaming period, were undergoing erosion, not deposition. From this area the whole of the Timiskaming is missing. Erosion of the mountains during Timiskaming time yielded a vast bulk of sediments to the lowland belt farther north, but it was not until they had attained great thickness and were in turn folded into the Algoman Mountains, and the old Laurentian Mountains south of the present Lake Superior had been reduced to flatness, that deposition was resumed in this southern area in Lower Hurionian times. Near Sudbury, however, where neither the Laurentian nor Algoman granites are found and where presumably neither mountain system was strongly developed, both the Timiskaming[11] and Lower Huronian Bruce systems were laid down. This portion of Ontario, therefore, exhibits a fuller stratigraphic column for these times than

[10] H. C. Cooke, "Land and Sea on the Canadian Shield in Pre-Cambrian Time," *Amer. Jour. Sci.*, Vol. XXVI (1933), p. 437.

[11] Assuming that the Sudbury series is equivalent, in part at least, to the Timiskaming.

the south side of Lake Superior, and consequently it has afforded better opportunity for piecing together the full history.

According to present views Lower Huronian rocks have not been found in the strip north of the Penokean trough. They should not be if that was the Algoman mountainous area undergoing erosion at that time. Twenty years ago the Ogishke–Knife Lake series on the northwest side of Lake Superior was supposed to represent Lower–Middle Huronian continental deposition while marine deposits were accumulating farther south in Wisconsin and Michigan. The Knife Lake series was followed in northern Minnesota by the great iron-bearing series, then classed as Upper Huronian. The striking unconformity between the two Van Hise and Leith recognized as one of the most conspicuous in the Lake Superior region.[12] Now we see why this break was so great. This was the area of the Algoman Mountains. The Knife Lake series and older rocks had there been deformed and intruded by batholiths. In Lower Huronian time the area was mountainous; erosion was in full sway, but deposition was elsewhere. The highlands contributed rock material to the Penokean trough south of Lake Superior. Gradually the highland area was brought low by denudation, and in Middle Huronian time, according to the present view, portions of it at least again became sites of deposition. In northern Minnesota the Middle Huronian Pokegama quartzite and iron formation (Gunflint, Biwabik) rest upon the much older Algoman plutonics and much-deformed pre-Algoman rocks, with the Lower Huronian absent.

Geologists working in eastern Ontario and Quebec describe the unconformity between the Timiskaming and the Huronians as a more striking structural break than that between the pre-Timiskaming and the Timiskaming. One reason for this would seem to be that the Canadian districts most studied in recent years are in the belt of the Algoman orogeny north of the original Laurentian area. In these northern districts Timiskaming sedimentation in many places followed pre-Timiskaming sediments and lava beds without strong intervening diastrophism, for the main Laurentian orogeny occurred elsewhere. In these places structural discordance between the older

[12] C. R. Van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *U.S. Geol. Surv. Mono.* 52 (1911), p. 604.

rocks and the Timiskaming, therefore, is not great. But, following the Timiskaming, the Algoman revolution strongly deformed the formations up to and including the Timiskaming, with the further complication of large intrusions, and a long denudation interval ensued before deposition was resumed in Huronian times. Consequently the unconformity between the Timiskaming and the overlying beds is structurally more striking in the Timiskaming belt than that between the Timiskaming and the rocks beneath it. For comparative purposes one should note that, between the time of the Laurentian revolution and the deposition of the earliest Huronian, there developed in eastern Ontario the post-Laurentian–pre-Timiskaming unconformity, the whole Timiskaming system, the Algoman revolution, and the post-Algoman–pre-Huronian unconformity, whereas on the south shore of Lake Superior this long sequence of important events is represented only by a single unconformity, between the Keewatin-Laurentian complex and the Lower Huronian. The very great difference in the histories of these two separate structural provinces has seemingly been the chief reason why the geologists studying one province were unable for so many years to bring their views into agreement with those working in the other.

In general the areas of Middle and Upper Huronian deposition were more extensive than those of the Lower Huronian. In the Middle Huronian period the Penokean trough widened to reach considerably farther south in Wisconsin as well as farther northeastward past Lake Timiskaming in Quebec.[13] In Upper Huronian time the sea encroached still farther over the old Algoman tract north of Lake Superior, but may not have progressed much beyond the Lake Superior region.[14] The Algoman Mountains had had their day and were now largely gone. Streams with reduced gradients brought to the sea large volumes of mud to be distributed widely as marine sediments, more locally as deltas, and perhaps also to be scattered on alluvial plains before reaching the sea (Michigamme, Tyler, Virginia, and Rove slates). Within the long span of the three Huronian periods foldings and uplifts occurred, locally at least, with important erosional and depositional consequences.

[13] Cooke, *op. cit.*, Map IV, p. 457.

[14] Van Hise and Leith, *op. cit.*, p. 613.

East of Lake Superior and north of Lake Huron the Huronian history is similar in some respects and different in others. The Bruce system is very thick and much of it is coarse, clastic material. Its thickening southward has been thought by some to indicate that the source of the sediments lay in that direction, but an alternative hypothesis is that the thicker portion of the Bruce may represent a deeper portion of the Penokean trough where greater sinking permitted greater accumulation of sediment than nearer the northern margin where the subsidence was less. Quirke and Collins prefer the latter interpretation, believing that the Bruce sediments came from the north.[15] The Cobalt and Whitewater systems are generally correlated with the Middle and Upper Huronians, though the Cobalt includes the glacial Gowganda formation, not recognized in the Lake Superior Middle Huronian, and iron formations are not conspicuous in either the Cobalt or the Whitewater.

The Keweenawan rocks occur in an area, somewhat elongate east and west, which centers in Lake Superior. The middle portion of this area, in Keweenawan times, was a sinking basin in which accumulated an immense bulk of lava flows and clastic sediments under continental conditions.

THE KILLARNEY GRANITES

The Killarney granitic batholiths were intruded during the Killarnean orogeny, either late in, or at the close of, the Proterozoic era. The time of this third great pre-Cambrian revolution affecting the southern portion of the Canadian Shield has not yet been altogether satisfactorily determined, but it is generally believed to have been after the emplacement of much of the Keweenawan system. Where are the Killarney granites? Unlike the Algoman granites, they are dominantly in a southern belt. They are where the Huronian formations had reached great thickness. They and the Killarnean folds rose along the site of the Penokean trough which had been to the south of the older folded Algoman ranges.

The northern limit of the Killarnean granitic intrusions so far recognized runs from near Staples in central Minnesota slightly north

[15] T. T. Quirke and W. H. Collins, "The Disappearance of the Huronian," *Geol. Surv. Canada Mem. 160* (1930), p. 3.

of east toward Duluth. Northwest of Lake Superior, between Duluth and Pigeon Point, are some narrow strips of Killarney granite, structurally related apparently to the Duluth gabbro and curving around to a nearly east-west trend as they point toward Lake Superior in the vicinity of Pigeon Point. They would seem to show more close genetic relationship to the Duluth lopolith than to the Killarnean folded ranges. The same may perhaps be true also of some of the granites mapped as Killarnean southwest of Duluth.

South of Lake Superior Killarney granite appears in many places over a broad belt from the lake shore southward to Stevens Point, Wisconsin, and apparently considerably beyond.

At the easternmost tip of Lake Superior Killarney granite appears on the Lake Superior map[16] in a narrow strip between Batchawana Bay and the Sault. Farther east it also occurs north of Lake Huron in small patches in the Blind River and Panache map areas.[17] Near Killarney, on Georgian Bay, begins an enormous area of Killarney granite widening steadily eastward. Its northwestern border cuts off the folded Huronian formations abruptly and runs northeast from Killarney, passing closely southeast of Sudbury and Lake Timagami, to the southern end of Lake Timiskaming and onward into Quebec. The belt of Killarnean folding is naturally wider than the batholithic belt. As diagramed by Collins[18] the border of the folded belt extends from north of Sault Ste Marie nearly due east for about 100 miles and then curves northward in a salient around Sudbury. Within the 100-mile east-west strip north of the North Channel of Lake Huron the folding dies out northward about 40 miles from the lake shore. Between Killarney and Sudbury the deformation of the Huronians has been more intense and the visible folded belt is wider than farther west. The southern limit of the Killarnean diastrophism is not known but must lie far to the southeast.

The Killarney Mountains thus developed in a southern belt which had received sediments from the more northerly Algoman Mountains of earlier generation. Their general position was not very dif-

[16] Leith, Lund, and Leith, *op. cit.*

[17] W. H. Collins, "North Shore of Lake Huron," *Geol. Surv. Canada Mem. 143* (1925), Fig. 8, p. 102.

[18] *Ibid.*, p. 102.

ferent from that of the main chain of Laurentian Mountains, though the latter may have been flanked by more outlying ranges. The later Killarnean deformation has greatly obscured the record of the earlier Laurentian orogeny.

In the western portion of the Lake Superior region the large Duluth gabbro lopolith, with accompanying intrusion of more acid magmas, complicates the picture. Just what relation it bears to the Killarnean folding awaits further study. Somewhat similarly placed on the north flank of the Killarnean disturbed area is the Sudbury norite sheet, also basin-like in structure.

GENERAL RELATIONS

The larger picture of the pre-Cambrian history of this portion of the Canadian Shield is therefore one of shifting belts of orogeny and shifting tracts of heavy sedimentation. The general regional behavior has not been very different from that of various other portions of the globe which, in later and better-recorded ages, have passed through several generations of mountain building involving migrations of geosynclines and belts of folding.

Neglecting the very early history of the region, the mountains of the first generation were the Laurentians, comprising a strongly developed southern system, and apparently also other prominent belts of folding and batholithic intrusion north of Lake Superior. In the denudation and eventual peneplanation of these, great volumes of sediment were carried onto the low tracts to the north of the southern system, giving rise to the Timiskaming, and from other more northerly ranges into the intermontane belts of the time, probably also contributing to the Timiskaming and producing in other areas various correlative formations. During the long Timiskaming period the Laurentian areas may have been affected by episodes of uplifting; in general they were areas of active erosion, not of deposition, though with the progress of time some of their marginal portions, particularly those of rocks less resistant than the batholithic granites in the heart of the foldings (and perhaps less strongly bowed up in later rejuvenations than the medial portions), may have been sufficiently reduced by erosion to receive later Timiskaming sediments.

The mountains of the second generation were the Algoman chains developed north of the earlier southern Laurentians in the region of thick Timiskaming sedimentation. Southward-flowing streams from the northerly Algoman Mountains brought detritus from their destruction back again onto portions of the old Laurentian area now very low. A depressed strip of this, including also a limited area around Sudbury which apparently lay between the southern Laurentian and Algoman mountain belts, became the Penokean trough in which Lower Huronian sediments accumulated in notable volume. During Lower Huronian time the Algoman Mountains were supplying sediments instead of receiving them. The Huronian periods were long, and in time the Algoman Mountains became reduced. Like later folded ranges they were probably rejuvenated from time to time, in whole or in part. Middle and Upper Huronian deposition took place over portions of the folded Algoman area as these became low. Keweenawan lavas and continental deposition followed in the Lake Superior region.

The mountains of the third generation were the Killarnean ranges developed south of the Algoman Mountains in the region of thick Huronian sedimentation. They in turn were subject to erosion which unroofed the granite batholiths. But since the Proterozoic this portion of the Canadian Shield has been a relatively stable area. No geosynclinal trough has formed and no great accumulation of sediments has taken place there. After peneplanation of the Killarney ranges, shallow Paleozoic seas spread from time to time over large portions of the low-lying shield as evidenced by remnants of Paleozoic beds which have retained their original flat-lying attitude but little modified until the present time. The Paleozoic geosynclines and Paleozoic mountains were elsewhere.

With this general conception in mind it is hoped that the student may escape some of the confusion which has commonly resulted from reading the conflicting views of different authors based on different portions of the Canadian Shield. The diversity of opinions expressed in the voluminous literature has been very great. It is hoped also that the history of any particular district which the student is studying may be better appreciated by a realization of the part which that area seems to have played in the general history of

the whole region. Probably also it will be easier to recognize and understand the especially significant features of the various portions of the shield if the student considers them in their relation to the general synthesis.

Certain interpretations have been followed in this study as if they were definitely established, whereas they are still open to debate. Additional facts may prove them to be erroneous. The tentative nature of the conclusions reached in this venture into a field of very difficult and complicated problems is to be kept to the fore. Doubtless some of these simple generalizations may require considerable modification to make them closely approximate even the larger outlines of the actual history; the details, of course, are largely lacking.

No attempt has been made to bring the Grenville sediments into the picture. For this will be needed more knowledge than we now have of the age of the intrusives which cut the Grenville. Are they Laurentian or Killarnean or of some other age? Radioactive determinations of the age of these intrusives are greatly needed. Reliable radioactive age determinations will likewise be of great assistance in classifying granites of unknown or doubtful relationship throughout the entire region under consideration. If they should reveal more episodes of granitic intrusion than the three here considered, they will but help fill out the fuller picture.

REVIEWS

Peary. By WILLIAM HERBERT HOBBS. New York: Macmillan Co., 1936. Pp. 502; maps 27; halftones and drawings 49; records and diagrams 10. $5.00.

It is fortunate that this life of Robert E. Peary has been written by one who knew Peary intimately, who is himself familiar with the Greenland ice cap, and is in other respects well fitted to portray the highlights and the shadows of a remarkable career which properly should have redounded greatly to the credit of the American nation, but which unfortunately has been poorly appreciated. The story is a thrilling one of solid achievement and extraordinary persistence in the face of great odds, told in a direct, distinctly literary style, following Peary's own style of plain fact without overdressing. The facts themselves are so striking and full of interest that their effect would only be marred for the discriminating reader by any attempt to play them up.

Part I, "The Period of Training," is relatively brief. Part II, "Exploratory Surveys," portrays the reconnaissance survey of the proposed Nicaragua canal route in 1885, Peary's first Greenland expedition in 1886, and his detailed work on the Nicaragua route in 1888. Part III, "Exploring the Arctic," covers the eventful years 1891–97. Outstanding was Peary's crossing of the North Greenland ice cap with his discovery of Independence Bay (now called Independence Fjord) on July 4, 1892. This hard-won success led next to his two most discouraging defeats in the sledge journeys of 1894 and 1895, when the odds against him were too overwhelming. That Peary and his companions ever got back alive on the 1895 journey was due only to the extraordinary character and competency of the leader and the fitness of his comrades. It will do anyone good to read these chapters; they are needed also to give the proper perspective to his later achievements when ripened experience enabled him to avoid certain serious difficulties and to overcome others. The two last chapters of Part III, on the recovery of the large Cape York meteorites and the very important discovery of the peculiar wind system of the Greenland ice cap, relieve the tension of the narrative.

Part IV is entitled "Direct Assaults upon the Pole," but it includes a great deal more than the title would suggest. In fact, although conquest

of the North Pole had been Peary's ultimate objective for a decade, his first direct attack did not take place until 1902, the last year of the "four-year expedition" (1898–1902). In 1898 and 1899 his sledge journeys resulted in the mapping of much of the largely unknown fjord region west of Kane basin, and in 1900, passing Lockwood's farthest north, he rounded the northernmost point of Greenland and explored some distance down the east coast. After two more winters in the Arctic, Peary set out for the Pole itself in the spring of 1902 but got only as far as 84°17′ on the ice of the Arctic Ocean.

As early as 1895 Nansen, drifting across a part of the Arctic Ocean in the "Fram" frozen tight in the moving ice pack, had attained 86°14′ N. with much less effort than Peary had been making, and in 1901 the Duke of the Abruzzi's expedition, likewise by boat in the ice drift, got as far as 86°34′ N. Peary was accordingly in the background, and for a time it seemed as though his plan of reaching the North Pole by dog sled from Greenland or Grant Land, to which he had devoted so many years, was not the most promising one. On his seventh expedition north in 1906, however, he vindicated his choice and methods when he pushed "farthest north" on to 87°6′. He might have gone farther but for an open lane of water, the "Big Lead," which had caused a delay of six days until new ice permitted crossing, and for a fearful storm later which had kept the party in camp for another six days while provisions were reduced and the shifting ice pack carried them 70 miles to the east. Three years later Peary finally accomplished his life-ambition by reaching the North Pole through perfect planning, skilful use of many supporting parties which enabled him to start the final dash from 87°47′ N., and, as it happened, favorable conditions of weather and ice.

Part V is "The Great Controversy." In more than twenty years of Arctic exploration, Peary had learned how to overcome successfully the natural difficulties of travel in the far north and had achieved his main objectives, but he did not understand so well the psychology of the general public, which had very little understanding of the Arctic. Nor did they understand the plain, blunt, rugged integrity of the man. It was his great misfortune that an obstinately gullible, public either could not or would not distinguish between the genuine and the false and showered partisan abuse upon him through many of the remaining years of his life. In a way it was our country's misfortune as well. This lamentable controversy Professor Hobbs has handled very fairly and judicially. He does not, for example, even mention Cook's long sojourn in the federal penitentiary at Leavenworth for use of the mails in oil frauds subsequent to the Mount McKinley and North Pole exploration frauds.

From Professor Hobbs's *Peary* the reader will obtain a vivid picture of some of the most notable and important Arctic explorations, with appreciation of the problems and a good understanding of the results. The volume is a valuable contribution to the geography of the Arctic region. Its several appendixes contain interesting information of various sorts, including Peary's instrumental readings made in the vicinity of the Pole and the independent computations by Hugh C. Mitchell and Charles R. Duvall of the United States Coast and Geodetic Survey, both of which give the latitude of Camp Jessup as 89°55′22″ N.

R. T. C.

Geologie von Spitzbergen, der Bäreninsel, des König Karl- und Franz-Joseph-Landes. By PROFESSOR DR. HANS FREBOLD. ("Geologie der Erde.") Berlin: Gebrüder Borntraeger, 1935. Pp. 195; figs. 81; maps 8.

The island groups described in this monograph lie near the outer margin of the great Barents Sea Shelf which is structurally a part of continental Europe. The major part of the work is devoted to detailed descriptions of the stratigraphy and structure of Spitzbergen and Bären Island. Less detailed accounts are given of the other islands. Because of the fact that early Paleozoic as well as the older rocks were involved in the intense diastrophism of Caledonian mountain-building and likewise in later deformations, the structure of these older rocks is extremely complex. The trends of Caledonian structures are direct continuations of those in the northern part of the British Isles. The general parallelism of Caledonian and Tertiary structures causes the various formations to outcrop in a series of belts extending in a north-south direction across Spitzbergen and Bären Island.

The oldest rocks are the Hekla Hoek, which attain a thickness of more than 10,000 meters and which include not only pre-Cambrian but likewise early Paleozoic. This great thickness comprises igneous and varied types of metamorphic rocks as well as sediments which are still difficult to differentiate. Deposition of clastic sediments between the great Caledonian folds and in gräben occurred in Devonian here even as in northern Britain. Some coal is found in Devonian strata on Bären Island but it is in Carboniferous rocks that it first reaches extensive development. The Upper Carboniferous and Permian are widespread and consist of much highly fossiliferous limestone in addition to extensive shales, sandstones, and some conglomerates.

The Triassic is of but local interest on Bären Island but reaches great dimensions on Spitzbergen, where it consists of both marine and terrestrial

sediments. The Jurassic is found on all the islands and represents more extensive transgressions of the sea than occurred in the Triassic, for the most important rocks are marine. The most widespread of all transgressions however, came in Cretaceous. Rocks of this age are found on all the islands, large and small, and in Spitzbergen they contain considerable quantities of coal.

Tertiary rocks are found only on Spitzbergen and, in addition to marine and terrestrial sediments of some variety, contain much coal.

Sills and dikes of dolerite and extrusions of basalt are found in Mesozoic and all older rocks.

The Quaternary is chiefly interesting for its glacial features. Pleistocene terraces which appear to have been developed during interglacial stages are found on all the islands. Moraines and striations indicate that the ice was formerly much more extensive than now.

L. M. GOULD

Physiography of the United States. By FREDERIC B. LOOMIS. Garden City, N.Y.: Doubleday, Doran & Co., 1937. Pp. viii+350; figs. 212; map 1. $2.75.

This book is intended for students with no previous training in physiography or geology. Ninety-five pages are devoted to general physiographic processes, rocks and structure, weather, soils, and a skeletal outline of geologic history. The remaining 255 pages form a well-balanced and in eresting description of our country adapted to the viewpoint of the beginner in the subject. The book should be a useful background for the student of human and economic geography.

The author's treatment of United States physiography has much to commend it. He emphasizes the description of salient, well-known, and characteristic features of each province rather than the tedious delineation of boundaries. His numerous sketches and photos, many of them drawn from familiar sources, are effectively employed. His own perspective diagrams are executed with a simple and rather crude technique. In the reviewer's opinion they are not only effective as illustrations but should stimulate in artistic students a desire to master and perfect this very useful method of depicting landforms. The author enlivens and deepens his subject by reviewing the geologic history of each province. And, best of all, the book is compact. Its brevity will permit the teacher to assign collateral readings in original sources, which the author makes no attempt to summarize.

The book is not without its weak points. There are not a few errors in the spelling of proper names. In certain instances both the references

cited and the point of view seem hardly in accord with our present state of knowledge. The reviewer grants that many concepts in physiography are controversial and that opinions differ; nevertheless, a college text should present fairly all sides of a subject or at least acknowledge recent studies in which opposing conclusions are reached. This the author fails to do in at least two major regions. He describes the Rockies in terms of a "Rocky Mountain peneplain" of Oligocene age, a "Prairie peneplain" of late Miocene or early Pliocene age, and an uplift of later date. No mention is made of the multiplicity of erosion levels recently proposed by Van Tuyl and Lovering, or of the "stripped surface" hypothesis applied by Loughlin in the Cripple Creek district, Bucher in the Beartooth Range, and others elsewhere to certain supposed erosion levels. Again, New England and the Appalachians are described almost wholly in terms of the Schooley and "lowland" peneplains together with Johnson's fall-line peneplain. No mention is made of Barrell's piedmont treppe, of the divergent views of such workers as Mrs. Knopf, Stose, Ashley, and Wright in the Southern Appalachians, or of Flint's hypothesis of glacial stagnation in New England. There are a few errors of fact, as when the author states that the Mosquito Range (Colorado) is a "low ridge east of the Sawatch Mountains . . . . high enough to prevent the Arkansas River from entering the [South] Park." Actually the Mosquito Range is one of the loftiest in the state.

Your reviewer regrets that the author generally did not see fit to assign concepts, ideas, and theories to their originators. Knowledge of United States physiography has developed gradually with many reconciliations of, or compromises between, somewhat opposing hypotheses. Granting that this volume is intended for the beginner, the reviewer nevertheless believes that appropriate reference to original sources and to the authors of outstanding concepts would enliven the subject.

The few faults of this volume must not be construed as limitations on its usefulness. The book is a competent treatment of its subject on a new level. As such, it fills a definite need and should meet with wide use.

W. E. POWERS

Introduction to Theoretical Seismology, Part I: *Geodynamics*. By J. B. MACELWANE, S. J. New York: John Wiley & Sons, 1936. Pp. x+366; figs. 67. $6.00.

This is the larger and principal volume of a two-volume work on theoretical seismology. Part II by F. W. Sohon, S. J., dealing with seismometry, appeared in 1932.

The present volume is an advanced technical treatise on seismology as

it relates to the earth and its properties. After an introductory chapter the author launches into a detailed and highly technical development of the theory of elasticity. This occupies the first three chapters. The two following chapters are devoted to developing the theory of elastic-wave propagation in the interior and on the surface of an isotropic elastic solid.

Chapter vi is an insertion consisting of a vector treatment by Sohon of the propagation of elastic waves, of which further mention will be made later.

Chapter vii treats of the reflection and refraction of elastic waves.

The foregoing chapters contain the essential theory with which to interpret seismic data. The chapters following are devoted to the methods of interpretation and the results obtained thereby. There is a chapter on the paths of seismic rays and the surfaces of discontinuity that have been discovered within the earth. This is followed by a chapter on the interpretation of seismograms, another on the determination of epicenters, and a concluding chapter on the depth of focus. In addition, there are appendixes and an Index.

Since seismology is still in a state of very rapid evolution, and since most of the general treatises on the subject in English are more than twenty years old and hence badly out of date, the present work fills an urgent need. It is up to date and gives copious references to the key literature of the various aspects of the subject. These are a great aid to the student and the research worker.

The book is written in the language of mathematical physics, which may be considered a handicap by some, but which is a fault of the subject matter and not of the author. No other way has yet been invented to analyze problems in elastic-wave transmission. The analytical treatment appears to be extraordinarily lucid.

The author is particularly to be congratulated in his discussion of the interior of the earth for stating his physical facts and sticking to them rather than of falling into the much too prevalent error of telling us what rock types occur at given depths. Unfortunately, seismological data do not give an answer to this question; neither do the data of geology.

On the negative side the book is not without its faults, the most grave of which, in the opinion of the reviewer, is the more or less inverted logical sequence in which the successive topics have been treated. As an experimental science seismology starts with observational data—earthquakes and seismograms. Theory is then developed to account for the data in terms of the physical properties of the earth. Deductions from theory then serve as guides to further observation. One would hardly suspect this, however, from the arrangement of the present treatise, which

is written rather in reverse. In fact, a physicist uninitiated in seismology would probably have considerable difficulty determining from this book just what are the experimental facts upon which the whole structure depends for its validity.

Another lesser fault is that the subject matter is somewhat loosely integrated. For example, considerable space is devoted to the theory of anisotropic elastic bodies, yet use is made only of the theory for isotropic bodies. A more flagrant case of the same kind is the insertion of a whole chapter devoted to the vector treatment of wave motion to derive what had already been proven by Cartesian methods. Since the book elsewhere employs the Cartesian geometry, and no further use is made of these vector derivations, it is by no means clear why they were included. They appear to serve no more useful purpose than to demonstrate that Father Sohon understands vector analysis.

In spite of these shortcomings, the book remains a welcome and valuable treatise of modern seismological theory.

M. KING HUBBERT

"Preliminary Report on the Gold Deposits of the Virginia Piedmont," by C. F. CLARK, JR., in *Virginia Geological Survey Bulletin 44*. Richmond, 1936. Pp. 42, figs. 14; pl. 9.

The field and laboratory work on which this report was based was done under the United States Geological Survey; the report, however, was published under the auspices of the Virginia Geological Survey. It is a report for economic geologists and those directly interested in gold workings in the area described. The mines which have been reopened are described in detail.

In the Virginia Piedmont ascending waters have deposited gold in a gangue made up chiefly of quartz, sericite, and ankerite, the gold occurring in intimate association with sulfides, notably pyrite. The deposits, in the Wissahickon formation, a schist generally considered to be pre-Cambrian, are either in the form of lodes or are replacement deposits. The occurrence of the ores is localized along tension cracks arising from a dominant thrusting force from the east.

The gold is recovered to a slight extent from placer mining but chiefly from its original occurrence in lode or replaced schist. Enrichment by specific-gravity settling is common near the surface, but in the deeper workings there has been no noticeable enrichment by solution and redeposition by the downward percolating waters.

JEAN GRACE